EVIDENCE
of
PRECAMBRIAN GLACIATION
in
NORTH AMERICA

Senior Thesis
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INTRODUCTION

The first 4000 million years of the earth's history falls within the Precambrian eon. The oldest preserved rocks, those of the Archaean Era, show no evidence of widespread glaciation. In Proterozoic strata, however, two broad periods of widespread glaciation are apparent. Evidence of the earliest known ice age, the Huronion, is found in Canadian Shield deposits dated at 2300 my. A younger, far more extensive period of glacial events is observed at the end of the Precambrian in North America and throughout the world. The major glacial episodes which followed in the Paleozoic Era - the late Ordovician and the Permo-Carboniferous ice ages - are poorly documented at best in North America. Possible glacial deposits of late Ordovician age have been reported in Newfoundland (McCann and Kennedy, 1974) and Nova Scotia (Schenk, 1972). The main focus of this report, therefore, will be the Precambrian glaciations in North America, as evidenced by tillite deposits and other glacial features, and their possible relationships to Precambrian glacial deposits elsewhere in the world.

EVIDENCE OF EARLY GLACIATION

Shortly after the theory of a Pleistocene ice age was introduced in the first half of the nineteenth century, speculation began on the existence of yet earlier glacial ages. Perhaps the earliest report of a Pre-Pleistocene glacial deposit was published by Selwyn in 1859, in which an apparently lithified till was found overlying a smooth and stratified surface of older rock. Of the many geologic features indicating ancient glaciation, this is the classical combination.

The most widely discussed of all relics of pre-Quaternary glaciation are tillites, defined by Harland et al (1966) as "unsorted, lithified coarse clastic rock of pre-Pleistocene age produced by glacial ice." The term "tillite" is a genetic one, refering to a deposit of glacial origin. The term "diamictite" is used for unsorted or poorly sorted terrigenous sedimentary rock with a wide range of particle sizes, without regard to mode of origin. Tillites are the most widespread of all glacial deposits, and are the most easily recognized. Tillites may be identified by the complete lack of grain sorting, the minerological immaturity of the components, and a massive to stratified structure with scattered outsize clasts. The grain sizes general range from clay to boulders, with a maximum boulder diameter of 5 m (Harland et al, 1966). The presence of rock flour in the matrix or extrabasinal minerals derived from distant sources are indicative of glacial origin of a diamictite. Faceted and striated stones and excessively grooved, polished and striated pavement are also highly suggestive of a glacigenic deposit. All of these feature require careful interpretation since most of them can be explained in more than one way. In sufficient quantity or in sufficient quality, however, these features may suggest a glacial origin.

It should be remembered that diamictites can form in a variety of ways, such as by mudflows or turbidity flows, so fabric analysis alone cannot be used as a criterion of glaciation. However, these and other forms of mass-movement are likely to be more common during glacial episodes. Thus, the presence of diamictites should alert the observer to the possibility of contemporaneous glaciation.

Another feature which may be diagnostic of ancient glacial deposits are rhythmically laminated sediments, often called "varved clays." Although often associated with tillites, they are absent in many glacial sequences. The rhythmically layered sediments of both glacial and nonglacial origins are termed "rhythmites." These may result from chemical or organic sedimentation as well as from mechanical deposition in a glacial lake. Most of these can be described as argillite or shaly siltstones, but they are characteristically arkosic if derived from crystalline rock due to the large content of unweathered plagiqclase and mica. The general character of stratification consists of couplets composed of a coarse-grained member, often a quartz silt, and a fine-grained clay member. Individual members may be laminated or graded, but the contact between the two successive couplets is sharp. These are apparently produced by turbidity currents originating from streams into glacial lakes, with the different members reflecting differences in discharge and sediment load during winter and summer.

The distinction between rhythmically laminated deposits of glacial and nonglacial origin must be carefully made. Nonglacial rhythmites, such as those of the Green River Formation of Wyoming, can be distinguished by the carbonate and organic content of its members. Also, the occurrence of marine fossils in apparent rhythmic deposits will preclude the possibility of a glaciolacustrine origin. Conversely, the presence of dropstones, especially if faceted or striated, is generally diagnostic of glacial sedimentation.

Stratified glacial deposits, either outwash or ice-contact stratified drift, are somewhat more difficult to identify than till and are present in smaller deposits. A varied arkosic lithology and clasts with glaciated shapes and striations are characteristic for both types of deposits. In ice-contact stratified drift, the fine fraction has been removed by meltwater, leaving the coarser sizes as unsorted lag gravels or well-sorted, well-stratified drift. The distinguishing features for these deposits are sharp lateral discontinuity of the units, abrupt changes in grain size both horizontally and vertically, inclusion of random lenses of diamictite, the presence of many poorly rounded clasts, and marked local deformation of beds. Outwash deposits in distal areas closely resemble coarse braided alluvial sediments of dry regions. To identify these as glaciogenic, characteristics such as the shapes and striations of the clasts, and the microtextures and shapes of sand grains have been examined. The presence of fresh plagioclase and feldspar in stream deposit sediments, and quartz sand grains so angular as to suggest crushing may suggest glacial origin.

Several minor features, involving structures and morphology as well as materials may be examined in older strata. Boulder pavements, which are flat-lying collections of boulders embedded in till with beveled and striated upper surfaces can be acceptable criteria for glaciation. These pavements are in turn overlain by more till. Stoss-and-lee features, asymmetrical blocks of bedrock, polished and striated on the upstream face and ice-plucked on the downstream f ace, are also very useful. In some areas, streamlined forms molded in the same tillite have been identified as drumlins. These forms run parallel to the dominant striations on the sub-till surface. Such features as these are also diagnostic. Finally, ice-thrust structures in stratified materials, formed by the drag of an overriding glacier, are commonly seen in Quaternary deposits, and should be present in earlier deposits as well (Flint, 1975).

LATE PROTEROZOIC GLACIATION

The glacial events of the late Proterozoic are recorded in three general regions in the North American continent. A series of diamictites are found in the western Cordillera from Alaska to New Mexico (sites 1 - 8 in Fig. A). Another group of glacigenic deposits and other features occur in areas of the Appalachian orogeny (sites 9 - 11), and a third group is seen in Greenland (sites 12 and 13). As the ages of the deposits range from 600 to 1000 Ma, these represent several episodes of glacial activity. Most of the diamictites are late Vendian (600 to 650 Ma) or late Riphaean (800 to 850 Ma) in age, but the tillites of the Death Valley region are apparently somewhat older (from 900 to 1200 Ma). Poorly-dated diamictites, falling anywhere from late Precambrian to Carboniferous, are found in Massachusetts, but there is no clear evidence of a glacial origin for these deposits, and they will not be discussed further.

Prior to this period of glacial activity was a lengthy interval (from 1000 to 2300 Ma) with little or no evidence of extensive glaciation in North America or elsewhere in the world. This hiatus remains difficult to explain, considering the periodicity of glacial events in the times that followed.

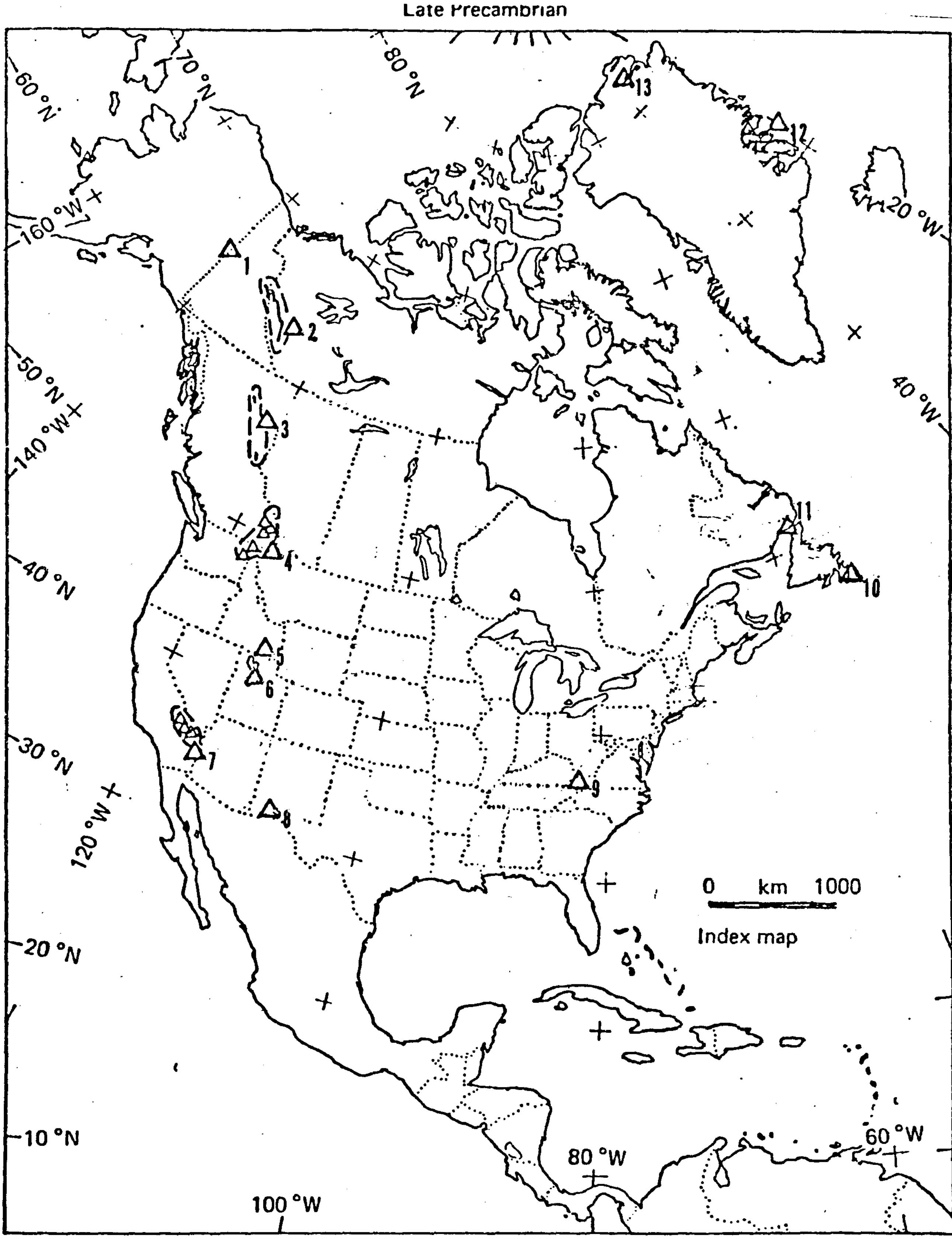

Fig. A Distribution of late Proterozoic glacigenic rocks of uncertain age in North America. Numbers refer to sections in text. (Modified from Hambrey and Harland, 1981).

1. East-Central Alaska

The upper part of the Tindir Group of east-central Alaska was first described by Mertie in 1933. Included in the lower part of the Upper Tindir Group are several unnamed formations of glacial origin. The best exposed section of glacigenic rocks is found along the north bank of the Yukon between 65°01' and 65°62'N latitude and 141°05' and 141°08'W longitude (Fig. 1-1). The Upper Tindir Group is preserved in a series of gentle folds with easterly-trending axes. Although the area has undergone considerable faulting, the rocks are not metamorphosed and primary structures are well preserved. While the exact age remains difficult to determine, available evidence suggests the glacigenic rocks are of the earliest Cambrian or very late Proterozoic age.

Evidence of glacial influence may be seen in units 2 and 3 of the Upper Tindir Group (Fig. 1-2). Coarse-grained interbeds of sandstone, siltstone and polymictic conglomerate are dispersed throughout the laminated red mudstones of unit 2, with proportionately more conglomeratic beds toward the top of the unit. Some sandstones are graded, and some mudstone beds are quite rich in hematite. Above this lie the thick-bedded diamictites of unit 3. The clasts (some of which are striated and faceted) are mostly dolostone with some chert, volcanic and mafic plutonic clasts as well. While the lower boundary of unit 2 appears conformable with underlying volcanics, an angular

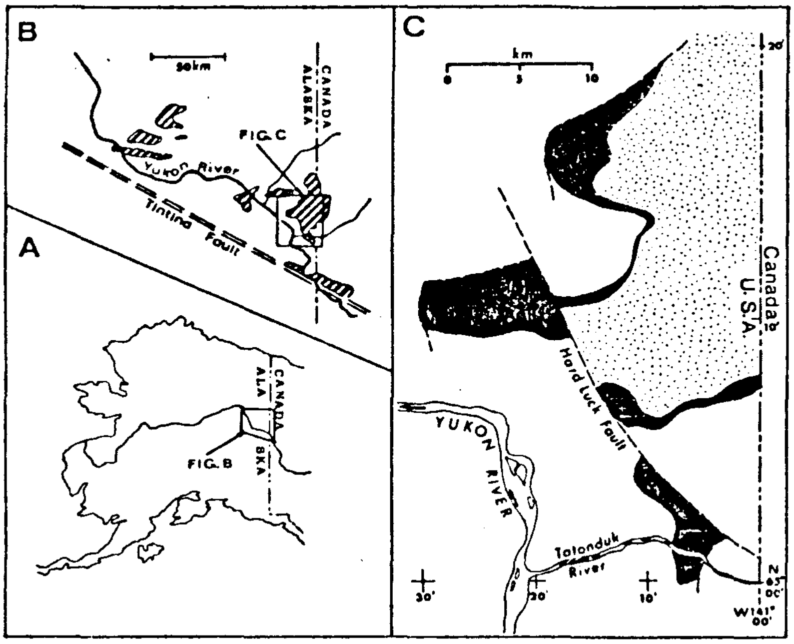

Fig. 1-1 Location map to show the distribution of glacigenic and associated rocks of the Upper Tindir Group in East Central Alaska. In Part B of the Figure, diagonal ornamentation shows the distribution of Tindir Group and related rocks. Part C shows (in black) the distribution of the 'basalt and redbed' unit as mapped by Brabb and Churkin (1969). Older rock units stippled.(Allison et al, 1981)

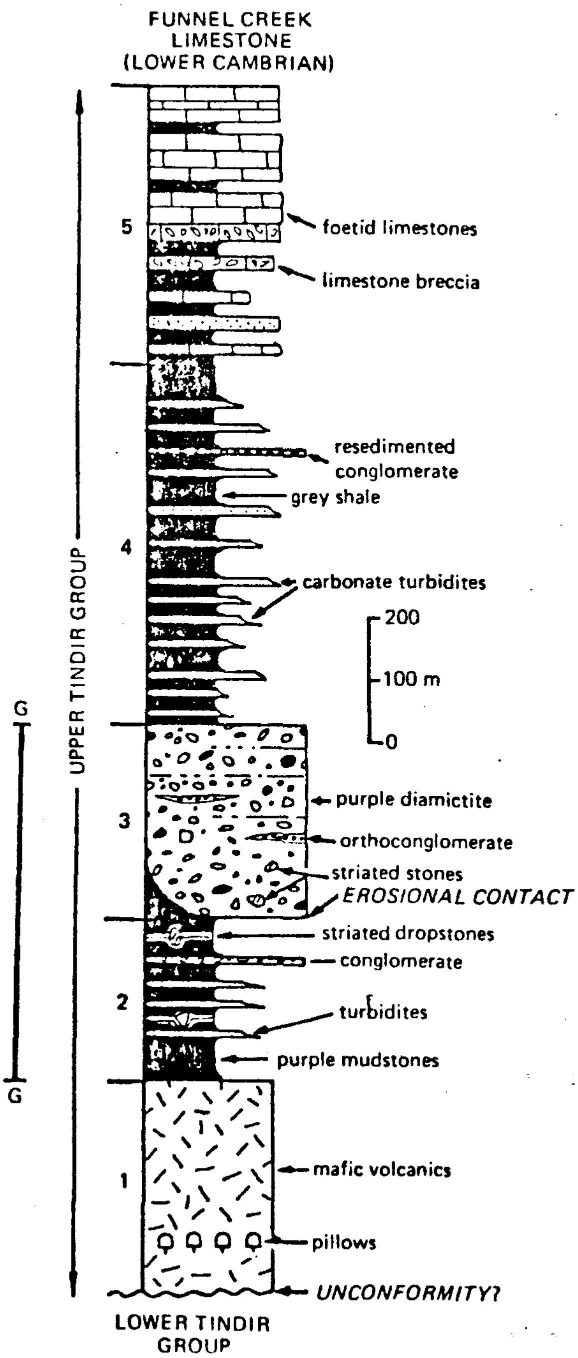

Fig. 1-2 Schematic representation of the stratigraphy of the Upper Tindir Group in the Tatonduk River area. Thicknesses shown are approximate. The section G-G is believed to be of glacial origin (Allison et al, 1981).

discontinuity appears between units 2 and 3 north of the Tatondak River. Formations overlying unit 3 are all conformable. The red mudstones of unit 2 are regularly bedded and laminated, with isolated striated clasts in some places. The siltstone and sandstone interbeds show graded bedding and occasional ripple cross-laminations show westerly transport. Unit 3 shows only crude bedding; the only layering present is observed in a few chert and jasper layers within the diamictite. The overlying shaly units are largely turbidite deposits. The wide range of clast size (granule to boulder), and clast shape (from rounded to angular), and the abundant striations and grooves cut on the surfaces of stones all suggest a glacial origin for these formations. (Allison et al, 1981).

The sedimentary rocks of unit 2 are interpreted as having been deposited below wave base, with sandstones and siltstones deposited by turbidity currents. Isolated clasts within laminated mudstones were apparently ice-rafted. Thin conglomeratic layers may have been produced by deposition from icebergs or sediment-laden ice sheets, or by mass flow deposits. The crudely bedded conglomerates of unit 3 are interpreted as glaciomarine. The presence of hematitic chert layers, occasional mudstone beds, bedded orthoconglomerates and stratification all support subaqueous deposition. Stratified clasts confirm the glacial origin of this unit. Both glacigenic units are interpreted as having formed in a marine basin into which glaciers were flowing from the east. The stratigraphy of the Upper Tinder Group closely corresponds to that of the Rapitan Group of the Yukon and Northwest Territories, suggesting similar conditions over the area. (Allison et al, 1981).

2. Northern Yukon-Northwest Territories, Canada

The Upper Proterozoic Windermere Supergroup of Northwestern Canada (Fig. 2-1) is a predominantly clastic succession between older Precambrian units and the base of the lower Cambrian quartzite. The base of the group is exposed in the Purcell Mountains, the Muskwa Ranges, the MacKenzie Mountains, and the Wernecke Mountains in the west-central portion of the Northwest Territories and the Northeastern Yukon areas (Fig. 2-2). The glacigenic Sayunei and Shezal Formations in the lower part of the group were deposited along a northwesterly-trending hinge zone of the early Cordilleran geosyncline. During deposition of the Shezal Formation, tectonic activity influenced the development of the basin. Despite the folding and thrust-faulting of the later Cordilleran orogeny, the primary sedimentary structures of the Proterozoic rocks remained intact. (Eisbacker, 1981).

The wide variations of facies and thickness makes generalization difficult. At the base of the group is the Sayunei Formation, ranging from 0 to 500 m in thickness, consisting of dark red siltstone-argillite couplets with

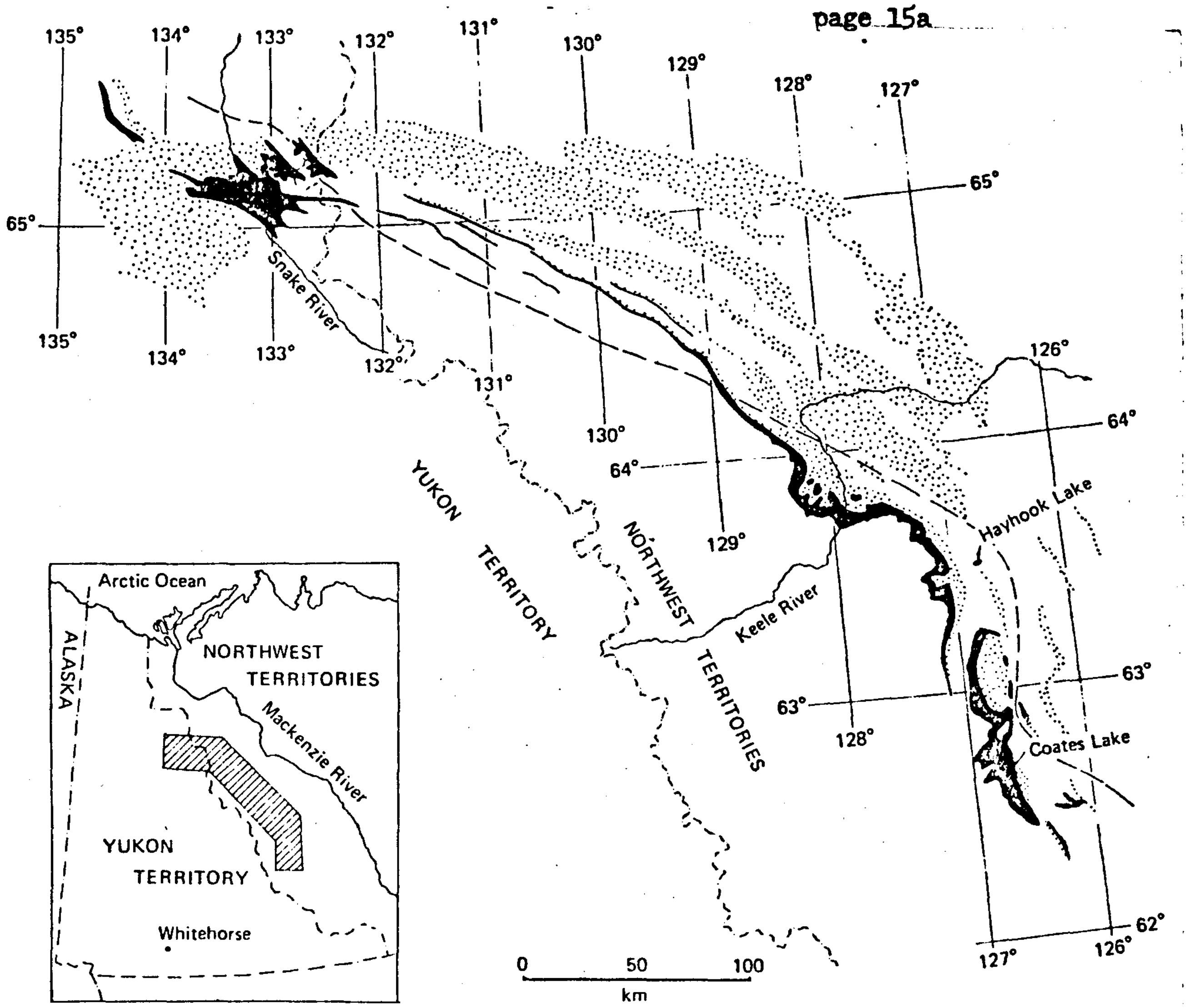

Fig. 2-1 Distribution of glacigenic rocks of the Rapitan Group (Windermere Supergroup) Older rocks; broken line; boundary between stratified facies(SW) and massive facies (NE). (Eisbacher,1981).

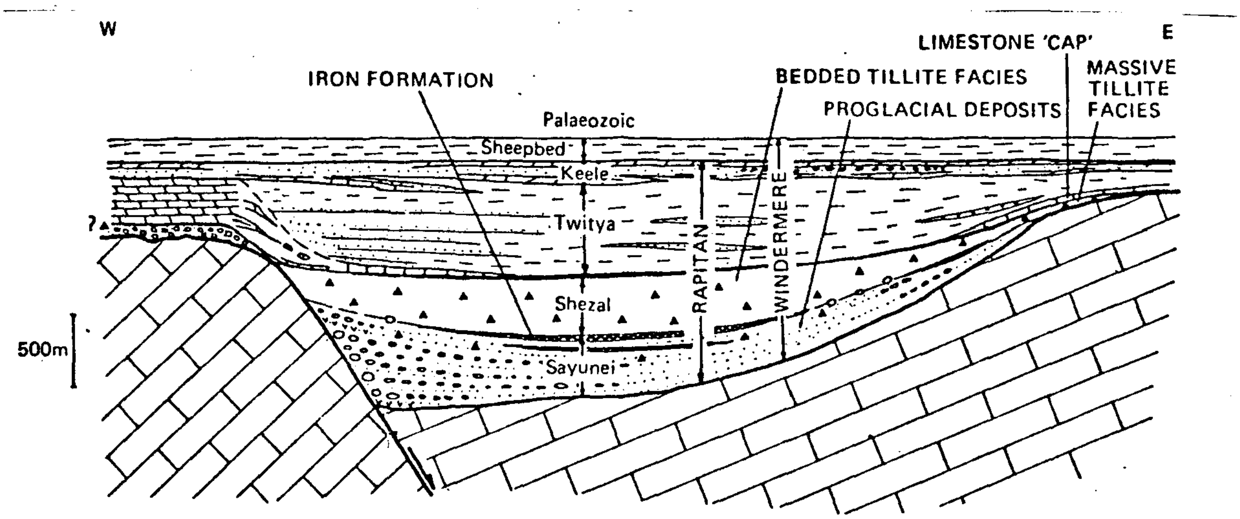

Fig. 2-2 Schematic W-E section through the Rapitan Basin slightly north of 65 N latitude showing stratigraphic position of the glacigenic Shezal Formation and the glacially influenced Sayunei Formation (Eisbacher, 1981).

dropstones, laminites, till pellets, and a jaspilite-hematite iron-formation member near the top of the sequence. Marine and nonmarine conglomerates represent the Sayunei Formation towards the northwest margin of the basin, where crudely stratified lenticular mass-flow deposits make up the formation. The overlying Shezal Formation contains a nonmarine tillite with numerous striated stones which lies locally on polished pavement, and a bouldery, glaciomarine diamictite laid down in a shallow marine environment. Internally unstratified tillite is seldom more than 10 m thick, but tillite sheets with vestiges of bedding can be up to 50 m thick. The massive, bouldery diamictite is characterized by an abundance of oversize, striated stones in a structureless sandy matrix. Cross-lamination of silty matrices can be seen in many diamictites despite the presence of numerous oversize stones. Between diamictic sheets, bedded clastics showing cross-bedding, cross-laminations and conyclote bedding can be observed. Although rarely found in the Shezal Formation, graded bedding is common in the Sayunei Formation. Dropstones are found in delicately bedded units of the basal formation. (Eisbacher, 1981b).

The numerous striated stones in the Shezal Formation along with the till pellets in siltstone laminites and dropstones in the iron-formation member of the Sayunei Formation strongly suggests a glacial origin for both units. The wide variations in facies and thickness point to a rapid transition from piedment ice to ice shelf and icebergs. Sharp regional variations in facies may also be the result of considerable relief near the shoreline (on the order of a few hundred meters). The deposits of the Windermere Supergroup as a whole have been interpreted as a clastic wedge along a rifted continental margin (Stewart, 1972). Other tillites occuring in this basin are the Toby Formation, the Mineral Fork Tillite and the Kingston Peak Formation. Strong similarities exist among these tillites of Australia and China, but whether this is a result of deposition within a single great basin later separated by sea-floor spreading, or a reflection of broad similarities in the environments of deposition remains unclear.

3. Northern British Columbia

In the Mount Lloyd George area of the Northern Rocky Mountains (Fig. 3-1), an unnamed succession containing Upper Proterozoic diamictites was recently described by Eisbacher (1981). The succession rests with sharp angular disconformity on a Middle (?) Proterozoic platform carbonate assemblage, which had been faulted, intruded by dikes, and beveled by erosion prior to deposition of the diamictites. Despite deformation from the late Mesozoic Cordilleran orogeny, sedimentary structures remain fairly well preserved in the glacigenic deposits.

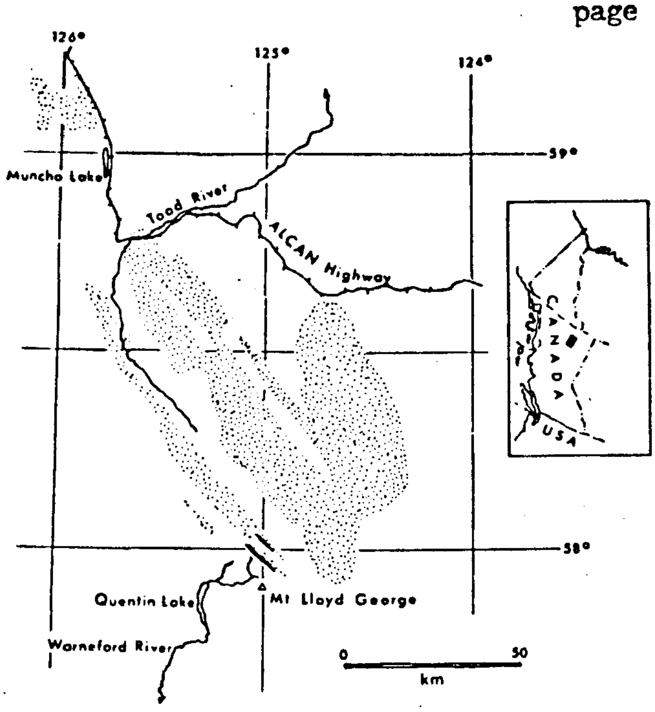

Fig 3-1 Location of the Mount Lloyd George Diamictites. Dotted pattern indicates older rocks (Eisbacher,1981).

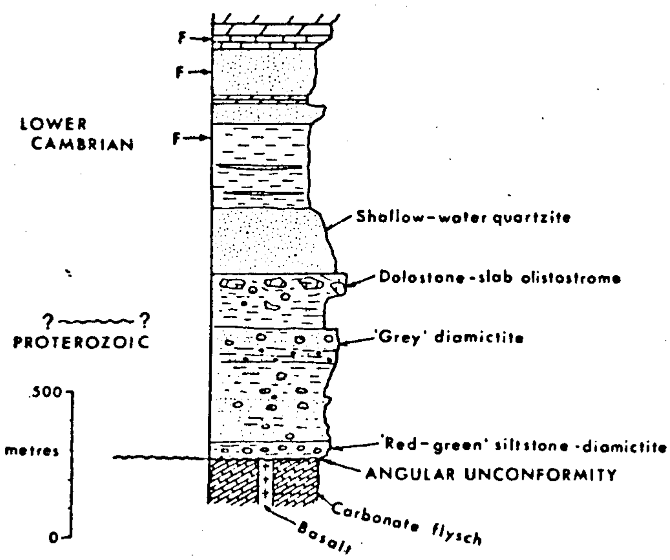

Fig. 3-2 Composite section of the Mount Lloyd George Diamictites. F, Fossil localities after Fritz (1972) (Eisbacher, 1981).

A generalized section of the 'Mount Lloyd Diamictites' begins with red-green diamictite, siltstone, sandstone and shale grading upward into gray shale and polymict diamictite (Fig 3-2). This is overlain by an olistostrome containing slabs of orange dolostone and cobbles of quartzite. Above this is a shallow water quartzite which grades upward into thinly bedded quartzite and silty argillite. The lower red-green interval consists of thinly bedded fine sandstones, and graded beds of granule-pebble conglomerate with oversize stones and rip-up clasts. Striated cobbles and boulders, some with distinct flat-iron shapes, are abundant. Stones are generally subrounded to rounded with dolomitic clasts being subangular. The diamictites have been dated at latest Proterozoic.

The Mount Lloyd George succession has been interpreted as a mass-flow along a steep submarine slope. The dropstone-like depressions associated with oversize stones in laminated beds and striated stone surfaces suggest possible glacial origin. Paleoslopes were apparently steep and irregular, controlled by faults as seen by the discontinuity of the beds. (Eisbacher, 1981).

4. British Columbia, Washington and Idaho

The Toby Formation is exposed in a sinuous homoclinal belt trending north-northeast from northeastern Washington to the northern Purcell Mountains in British Columbia, and in dislocated outcrops in the north central Purcell Range (Fig. 4-1). Metamorphism and structural overprinting from Mesozoic times have produced tectonite fabrics within the formation, characterized by flow and fracture cleavage, reorientation and deformation of clasts, and formation of small-scale deformation in schists. This unit overlies the metasedimentary rocks of the Purcell System and forms the basal unit of the Windermere system. Deposition of the Toby Formation follows Purcell sedimentation, which ended about 850 Ma, and predates extrusion of Windermere volcanics dated at 827-915 Ma (Miller et al, 1973).

The lack of orderly vertical or lateral stratigraphic variations of texture, composition, thickness and associations of lithology make generalization of the Toby Formation difficult. Typically, the basal beds are diamictite, the middle section is an interstratified group of diamictite, conglomerate, sandstone and argillite, and the upper portion is argillite and fine-grained, graded sandstone with dropstones. Thickness of the formation varies tremendously, from a few meters to 1800 m within an area of 10 km. The basal contact is disconformable or angular unconformable, demarked by sharp lithologic changes, while the upper contact is gradational into the argillites, sandstones and conglomerates of the Horsethief Creek Group. Volcanic agglomerates, breccias and conglomerates are interstratified with the uppermost diamictites in the southwestern-most exposures. In the basal Horsethief Creek

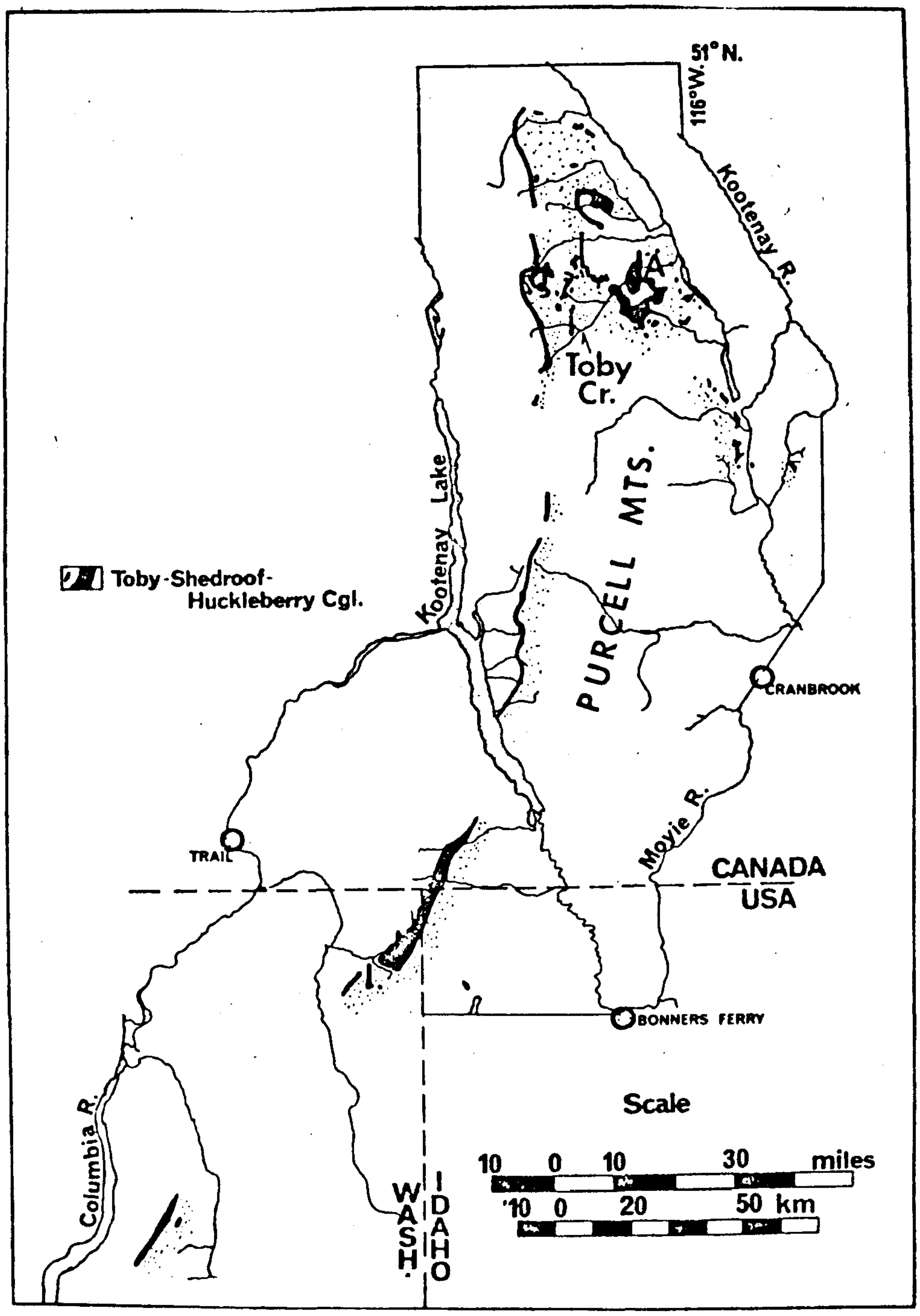

Fig. 4-1 Distribution of Precambrian tillite-bearing units of the Purcell mountains. Toby Formation shown in black; Purcell System adjacent to Toby Formation stippled. A indicates type section of Walker (1926). (Aalto,1981).

Group are diamictites lacking volcanic detritus overlying volcanic formations, suggesting that glacial conditions may have persisted throughout the entire period of vulcanism. The diamictite is generally gray unlaminated sandy mudstone with poorly sorted clasts of pebble to boulder size, enclosed in massive lenses or beds from 0.5 to 5 m thick. The clast-supported conglomerates have either an argillaceous or sandy matrix, and lack ordered internal primary fabric. Laminated or massive argillites are interstratified with all other lithologies or appear as lenses. Clasts within the diamictite are angular to well-rounded, and largely derived locally from the Purcell System (Aalto, 1981).

The composition, fabric, stratigraphic associations and lateral extent of the Toby Formation is best explained by glaciomarine deposition (Aalto, 1971). The unit was apparently laid down over an undulating land-surface. As continental glaciers pushed over the area to the sea, the bedrock was scoured and overlain by a grounded ice shelf. Initial basal melting produced poorly stratified till (the basal diamictite) which was partially reworked by bottom currents, producing thin lenses and interbeds of argillite, sandstone, mudstone and conglomerate. The uneven rates of ice advance over a landmass undergoing irregular isostatic and tectonic subsidence produced great lateral variations in thickness of accumulating sediment. As ice wasting continued, sediment gravity flows, ice-rafting and bottom currents produced the complex lithologies of the middle portion of the formation. With the retreat of the glaciers and a rise in sea level, sedimentation was dominated by turbidity currents, meltwater density currents and bottom current resedimentation, with some ice-rafting and submarine mudflows. Local submarine vulcanism produced intercalations of volcanic and glaciomarine rocks. Deposition of the Toby Formation ended with the complete deglaciation of the loclity, and was followed by pelagic/hemipelagic sedimentation and turbidity current deposition.

5. Southeastern Idaho

The Scott Mountain Member of the Pocatello Formation is a series of Upper Proterozoic diamictites in southeastern Idaho. The type area of the formation is the northern Bannock Range southeast of Pocatello, Idaho (Fig. 5-1), where it is exposed along a narrow north-trending belt 25 km long and 5 km wide. Incomplete sections are exposed along the faulted eastern side of the Oxford Mountains. This formation is the lowest rock unit of the late Proterozoic Cordilleran miogeocline (Stewart, 1972). Proterozoic rock units of the region have been metamorphosed to biotite-grade greenschist facies by the thrusting and crustal shortening of the Sevier orogeny. Clasts within the diamictite have been deformed, and primary sedimentary features have been obscured.

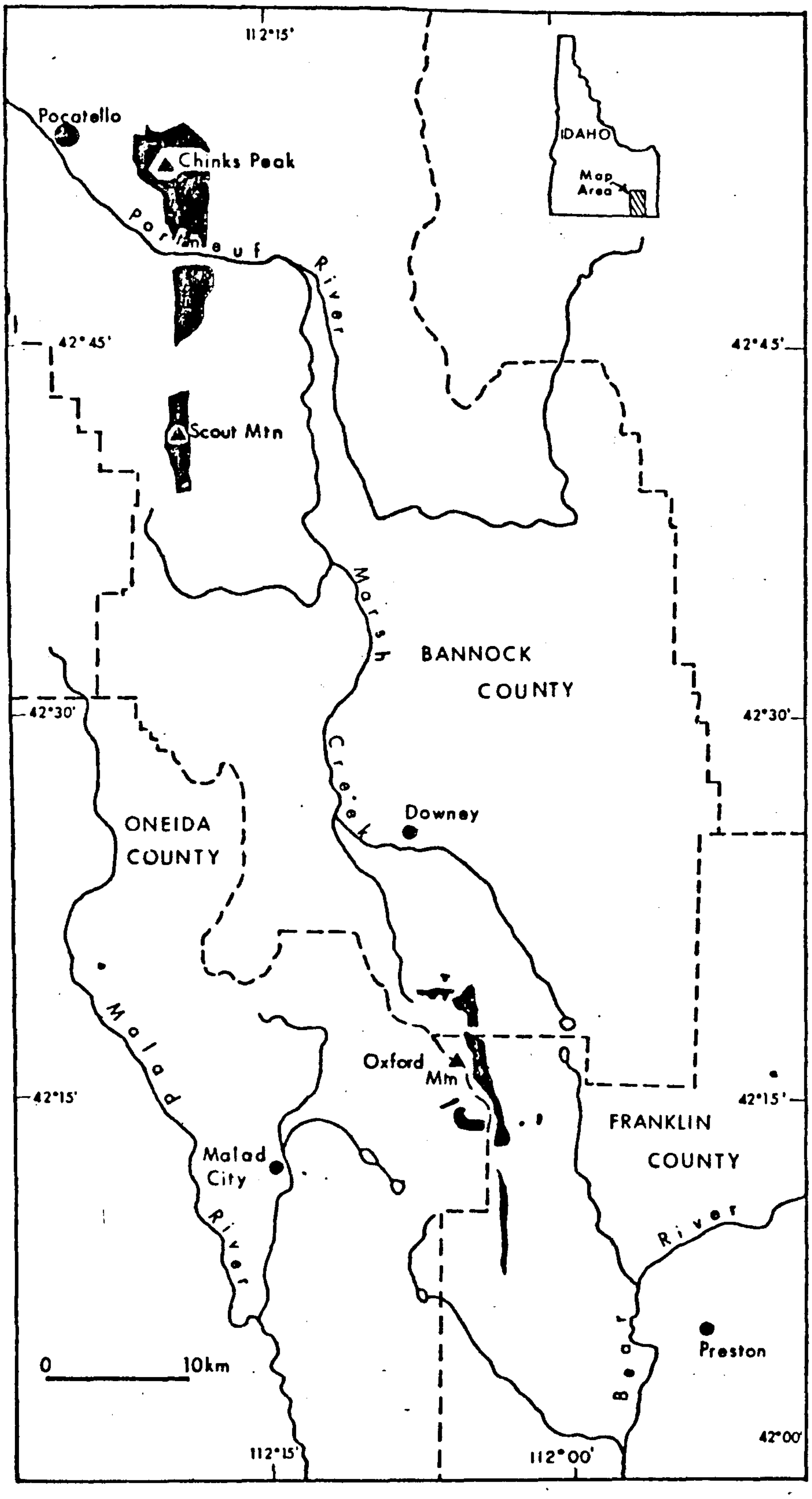

Fig. 5-1 Distribution of upper Proterozoic diamictites (Pocatello Formation) in southeastern Idaho. Diamictite outcrops in solid black (Link, 1981).

The type section, described by Crittenden et al (1971) and Trimble (1976) is taken at Pocatello, and is the most complete known. It contains a lower member of very thinly laminated dark gray to black silt and slaty argillite at least 250 m thick. Above this is the Scout Mountain Member, a sequence of interbedded diamictite, conglomerate, siltite, quartzite, limestone and dolomite somewhat less than 1650 m in thickness. A 20 m limestone bed overlies the lower conformable contact. Locally, this unit intertongues with the overlying Bannock Volcanic Member, a series of greenstones and pillow lavas exhibiting flow structure and pyroclastic texture. The uppermost member is mostly dark, thinly laminated argillites with quartzite interbeds over 700 m thick. Where visible, the upper and lower contacts of the glacigenic Scout Mountain Member are conformable. The diamictite is massive on a large scale and interbedded with fine-grained sedimentary and volcanic beds on a smaller scale (meters or centimeters). Sandy interbeds and conglomeratic intervals show graded bedding, cross-bedding, and contorted bedding locally. Although outsize stones up to 10 cm in diameter are found in thinly bedded sandstone, no unequivocal dropstones have been reported. A bed of pink dolomite less than 1 inch thick directly overlies a diamictite member near the Portneuf River. Finely laminated siltstones were reported by Ludlum (1942) who referred to them as "varved slates." Stones constitute up to 20% of the diamictite, and may be up to 1 m in diameter. The clasts are subangular to well-rounded, and composed of granite, gneiss, siltstone, quartzite and volcanics. While siltstones and volcanics are probably interbasinal, the other types were apparently supplied by an unknown source. Although in several localities stones have suffered deformation, reports of possibly striated but clearly faceted clasts have been made from Oxford Mountain (Crittenden et al, 1971).

A glacigenic origin for the diamictites was first proposed by Ludlum (1942). The Pocatello Formation itself is marine, as shown by pillow lavas and carbonate intercalations, as well as the lateral continuity of fine-grained clastics and stratified diamictite. The rapid thickness changes and penecontemporaneous vulcanism could suggest that tectonic activity influenced sedimentation. Stewart (1972) interpreted the paleogeographic setting to be a westward thickening miogeoclinal prism against the western edge of the North American craton. The deposits have been correlated on the basis lithology and age (estimated at 850 Ma) with those in Utah (Blick, 1981).

6. Utah

The Mineral Fork tillite of the Wasatch Mountains was first described and attributed to glacial processes by Hintze in 1913. Broadly equivalent diamictite-bearing units are found in the Dutch Peak and Horse Canyon Formations and in the lower members of the Sheeprock Formation of the Great Salt Lake area. Outcroppings of glacigenic beds are found in the Wasatch, Sheeprock, West Tintic Simpson and Deep Creek Mountains, and in a number of islands in the Great Salt Lake vicinity (Fig. 6-1). The Mineral Fork Formation is situated along an ancient hinge separating shelf sediments from miogeosynclinal sediments less than 870 Ma (Stewart, 1972). Bounded above and below by unconformities, the unit has been assigned a late Precambrian age, with the diamictites indirectly dated at about 800 Ma (Blick, 1981).

A generalized section of the Mineral Creek Formation taken from the Central Wasatch Mountain (Crittenden et al, 1952) consists of black diamictite, wacke, siltstone and shale with minor sandstone and conglomerate lenses overlying the quartzites and and varigated shales of the Big Cottonwood Formation. In other localities, the glacigenic unit appears as a green diamictite with conglomerate and greywacke, a black diamictite and mudstone, or a gray diamictite schist. The diamictite varies from well-bedded to massive, with local contortions of bedding. Lenticular laminites and rhythmites occur, but do not have the characteristics of varves (Condie, 1967). The diamictite is usually pebbly and moderately sparse in stones, but locally may have cobble- to boulder-sized stones in great abundance. Composition and shape of stones are highly variable. Ojakangas and Matsch (1976) reported dropstones in finely laminated beds and isolated outsize stones in fine-grain

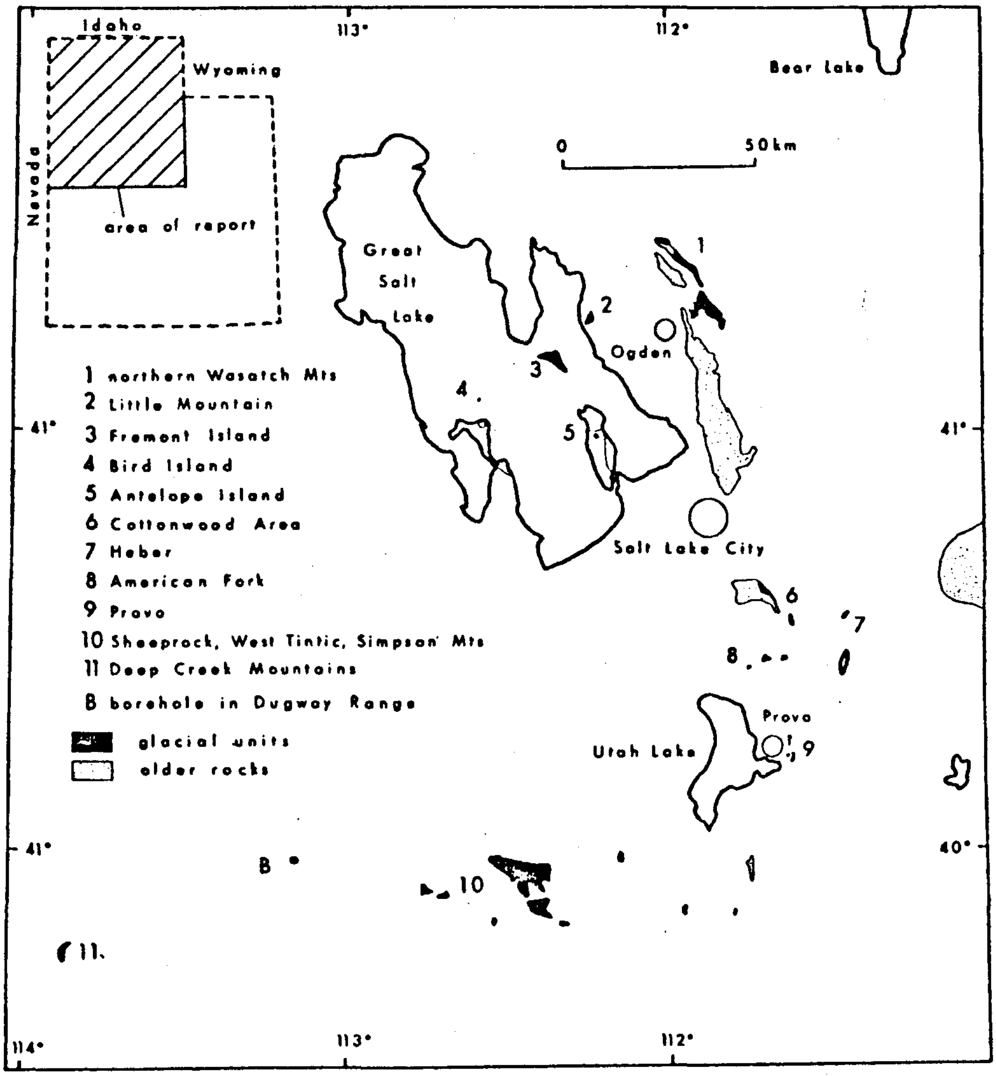

Fig.6-1 Distribution of Precambrian glacial units in Utah (Blick,1981).

beds.

The origin of the diamictites has been vigorously argued over the years. Condie (1967) suggested the formation was produced by tectonically-triggered slumping of glacial debris. Schermenhorn (1974) has expressed the view that the diamictites are not of glacial origin. Ojakangas and Matsch (1980) reaffirmed a glacial origin for the tillites, pointing to the presence of a polished, grooved striated bedrock surface at the basal contact of the unit. This was concluded not to be the result of thrust-faulting activity, but of glacial abrasion. Over the region as a whole, the thick sections of lenticular, stratified diamictite, the absence of varves and the presence of dropstones and rare pillow lavas suggest marine conditions were present. Erosional features, sedimentary structures and facies considerations point to a westerly direction of sediment transport (Blick, 1981). The tillites were apparently deposited in a miogeoclinal prism on the edge of a continental plate, according to Stewart (1972). Continental separation took place about 850 Ma ago, prior to glaciation. A chain of glacigenic units was deposited along this ancient hinge line which separated shelf sediments from miogeosynclinal sediments. This was later followed by crustal shortening in Cretaceous time, and crustal extension in the Cenozoic (Armstrong, 1968).

7. Death Valley, California

The Kingston Peak Formation of Death Valley, California, first described by Hewett (1940), is exposed in a discontinuous belt extending from the Funeral Mountains through the Panamint Range and southeastward (Fig. 7-1). The formation is the youngest member of the Pahrump Group, an upper Precambrian series which unconformably rests upon a metamorphic and igneous basement complex. The Pahrump Group is comprised of, in descending order, the Kingston Peak Formation, the Beck Spring Dolomite, and the Crystal Spring Formation. Although affected by late Mesozoic and Cenozoic tectonism, most exposures are relatively undeformed, and primary sedimentary features are well-preserved.

Northern and southern facies are distinguishable in the Kingston Peak Formation in the southeastern outcrops. The northern facies contains three persistant members. The lower member, ranging from 100 to 500 m, consists of fine quartz sandstones and argillite. The middle member, from 150 to 400 m, is entirely massive diamictite with outsize rock fragments. The uppermost member, from 1000 to 2000 m in thickness, is fine-grained sandstones and siltstones with conglomerate and diamictite interbeds. Each member is conformable at the base. The southern facies is comprised of a lower member of interbedded sandstone and argillite, with thin diamictite lens, and an upper member of thick diamictite, conglomerate, and conglomeratic sandstone with

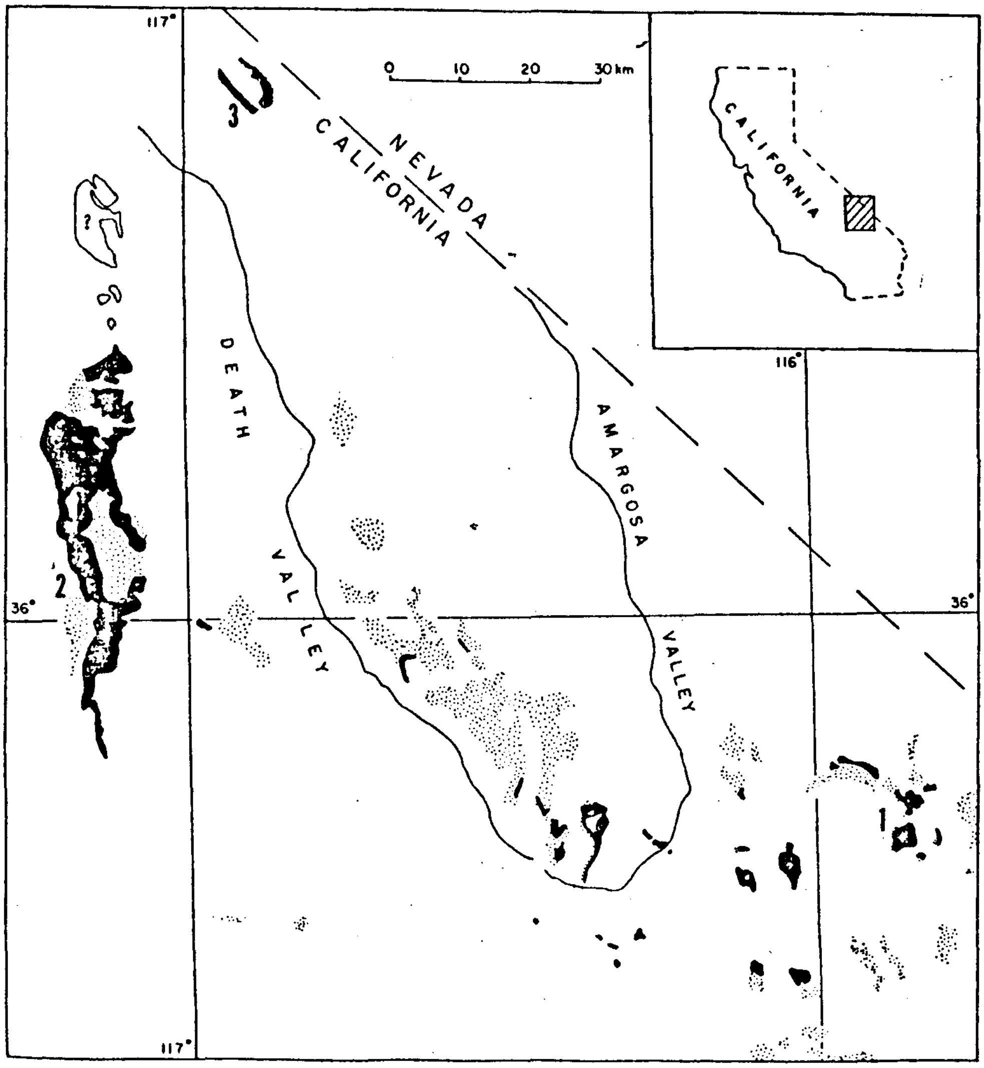

Fig. 7-1 Distribution of kingston Peak Formation (black) and older rocks, both Pahrump Group and basement complex (stippled), in the Death Valley region, California. Localities numbered as follows: 1, Kingston Range; 2, Panamint Range; 3, Funeral Mountains (Miller et al, 1981).

lesser amounts of argillite and sandstone. The lower member is bounded by an unconformity at its base, while the upper member is conformable with the lower member. The beds typically have marked variations in thickness, and lateral as well as vertical variations in lithology. The diamictite is massive, with rare sandstone interbeds. The conglomeratic sandstones, sandstones and shales of the upper Kingston Peak Formation are well-bedded, with cross-bedding, grading, and outsize stones in some localities. Graded sandstone, graded siltstone, and laminated siltstone with lonestones are found in at least one southeastern locality. Graded bedding with scour and fills, flute casts, convolute laminations and load structures are reported in some areas. The stones range from angular to well-rounded, comprising 10 - 20% of the diamictites. These stones, mostly less than 10 cm but ranging up to boulder size, are mostly intrabasinal quartzite, limestone, dolomite and gneiss, with a few pebbles showing striations or faceting (Miller, 1981).

A glacial origin for this diamictite is indicated, but the evidence is not conclusive. The process by which the unit formed may have been glacial deposition (Hazzard, 1939), submarine slides or debris flow (Johnson, 1957; Basse, 1978). Nonmarine conditions in the Kingstone Range are attested to by the presence of fanglomerates (Hewett, 1956) while subaqueous deposition continued elsewhere, as indicated by pillow lavas, well-developed turbidite, and ice-rafted dropstones. Deposition appears to have taken place along the eastern edge of the north-south trending Cordilleran geosyncline, shortly after rifting (Stewart, 1972). Wright, et al (1974) saw sedimentation taking place in a subsiding basin in the southeast trending portion of the belt. This would explain the more complex intertonguing of facies and distinct source directions in these outcrops. In the Panamint Range, the formation was laid down on a shelf to the west of the northwest-trending trough. Dolomite interbeds formed in the shallow waters in some sites, and offshore islands provided material for the diamictite.

8. New Mexico

An unnamed tillite unit occurring in the Florida Mountains of southwestern New Mexico was described by Corbitt and Woodward in 1973 (Fig. 8-1). The tillite unconformably overlies Precambrian mafic gneiss and is unconformably overlain by the Late Cambrian - Early Ordovician Bliss Formation, and no precise age has been determined for the unit, but it is assumed to be late Precambrian.

The tillite, as described in the Florida Mountains, is 12 m thick, consisting of a lower reddish, fissile shale member, a middle, greenish-gray shale, and an upper member of granitic boulders in an arkosic matrix. Enclosed in the two lower shales are rock clasts composed of sandstone,

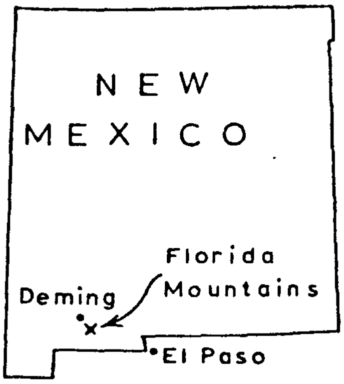

Fig. 8-1 Location of the late Precambrian tillite of the Florida Mountains, New Mexico (Corbett and Woodward, 1973).

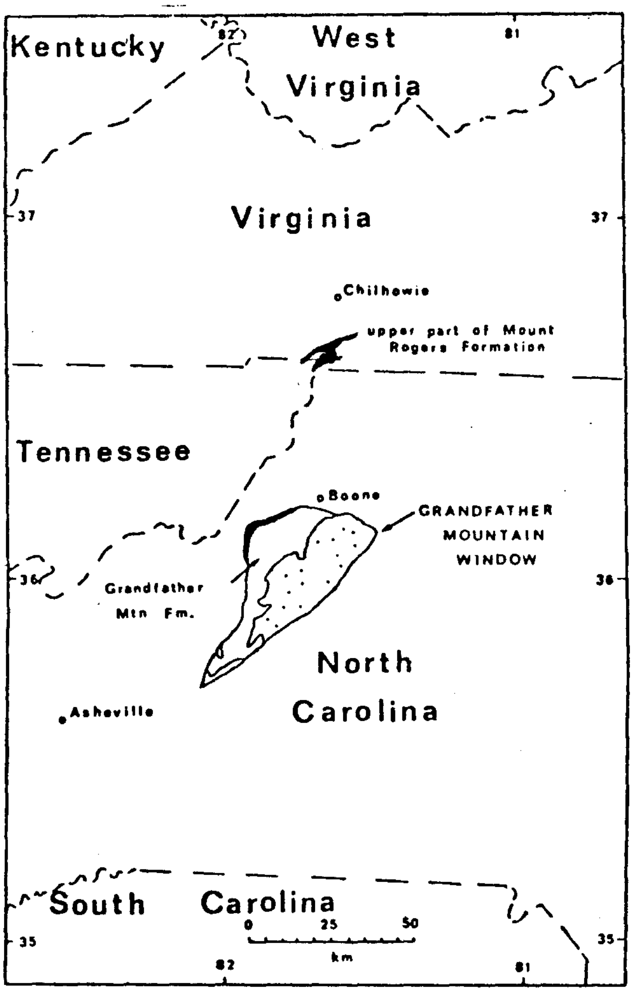

Fig. 9-1 Distribution of glacigenic rocks of the Mount Rogers Formation and Grandfather Mountain Formation. Older rocks of the Grandfather Mountain Formation are stippled (Schwab, 1981).

breccia, ironstone, diabase, basalt and granite. The clasts infrequently have striation and poorly faceted faces. Some of the boulders of the middle member appear to be dropstones, deforming the bedding within the underlying shale. Pellets of clay up to 2 mm in diameter are also present in the shale beds.

The lower shale members have been interpreted by Corbitt and Woodward to have been formed by pebbles, boulders and cobbles being dropped onto unconsolidated marine mud as ice-rafted clasts. The upper bouldery tillite was locally derived and deposited after a small area containing the dropstone-bearing shale was downfaulted. Local granitic debris filled the depression.

9. Central Appalachians

In the upper 1000 m of the Mount Rogers Formation of the central Appalachians, a late Precambrain tillite sequence has been reported (Jonas and Stose, 1939; Schwab, 1981). The best exposures are found in a 10 km by 60 km belt trending northeastward from the common corner of Tennessee, North Carolina and Virginia (Fig. 9-1). Analogous tillites are found in the Grandfather Mountain Formation, which outcrops in the Blue Ridge province of North Carolina. Although both formations have been affected by folding, faulting and low-grade metamorphism, the glacigenic units are well preserved and only slightly deformed. Both units are completely non-fossiliferous; the formations are conformably overlain by early Cambrian strata, and underlain by felsic volcanics, dated at 820 Ma (Rankin, et al, 1969).

The bouldery, unstratified diamictites of the Mount Rogers Formation occur as individual lensoidal members ranging from 1 to 50 m in thickness and in numbers from 1 to 10 (Fig. 9-2). The deep red rocks are commonly interbedded with rhythmic sequences of green siltstone and maroon shale (argillite), in which dropstones are usually found. Beneath the glacigenic facies lies a 50 to 150 m succession of fluvial arkosic sandstone and conglomer ate, while thicker alluvial deposits lie above the diamictite sequence. Individual diamictite beds lie conformably on the 'varved' dropstone-bearing argillites and shale, and are transitional to and conformable with overlying rhythmite sequences. The thick, structureless diamictite beds contain stones of sedimentary, metamorphic and igneous rocks in a hematite-rich mudstone matrix. These clasts are subangular to angular in shape, and mostly coarse cobbles to fine boulders, but sorting is very poor. The sequence of rhythmically alternating siltstone and maroon shale occur above and below the diamictites. The abundance of dropstones in some rhythmite horizons qualify them as laminated pebbly, cobbly and bouldery mudstones. The diamictite and dropstone horizons are virtually identical in the Grandfather Mountain Formation, but the laminated pebbly, cobbly and bouldery mudstones are less abundant

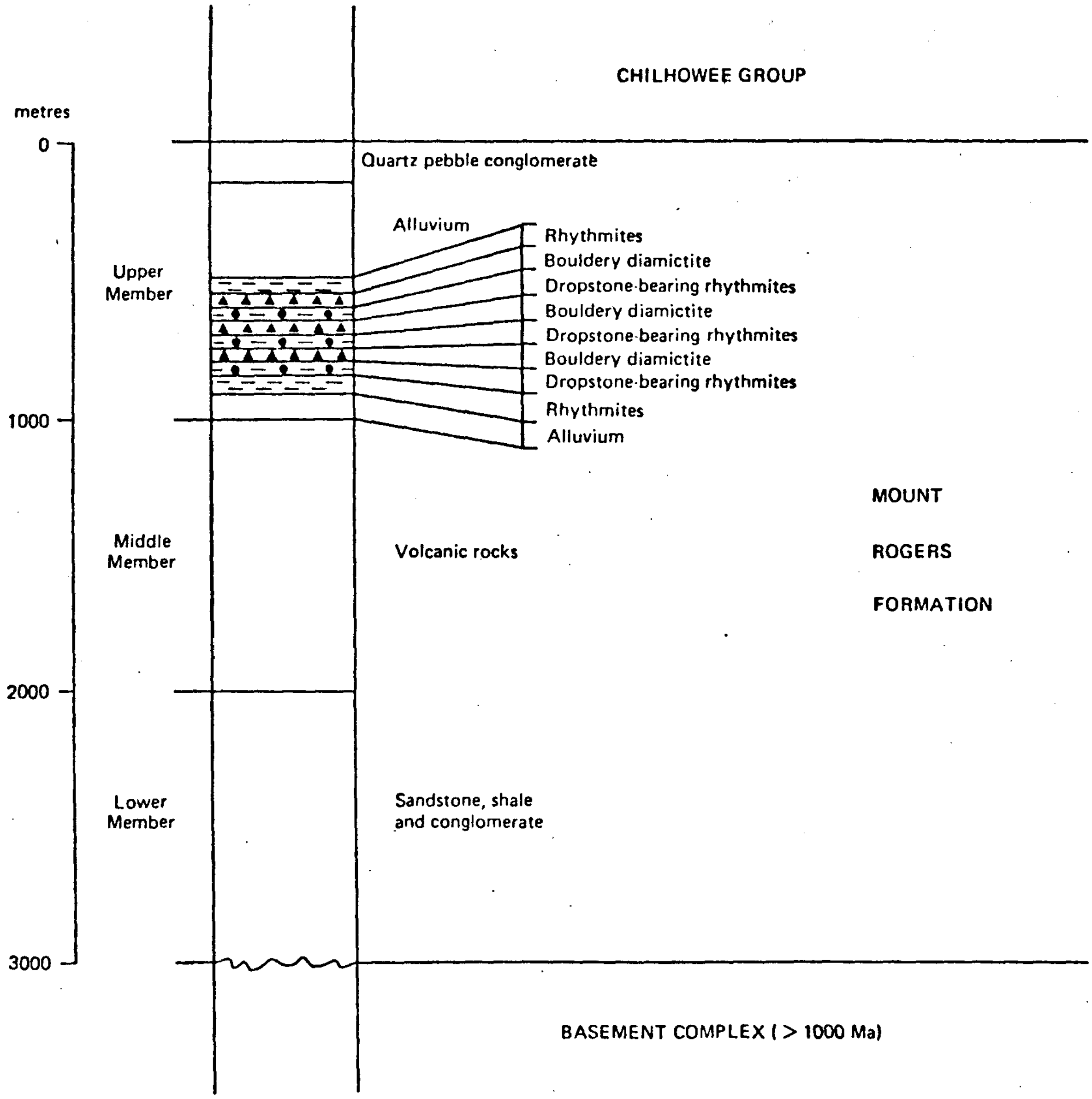

Fig. 9-2 Generalized stratigraphis section of the Mount Rogers Formation, which rests on Precambrian basement and is overlain by the Cambrian Chilhowee Group. The tillites of the Upper Member of the Mount Rogers Formation include both bouldery, unstratified diamictite as well as dropstone-bearing laminated pebbly cobbly and bouldery mudstone (rhythmites). (Schwab, 1981).

(Schwab, 1976).

Blondeau and Lowe (1972) interpreted the coarse bouldery diamictites of the Mount Rogers Formation as the ground moraine deposits laid down during repeated episodes of glacial advance and retreat. The rhythmites and dropstone rhythmites in both formations were probably formed below wave base in either a deep freshwater lake or marine basin, with dropstones being transported by icebergs or glacier-fed ice rafts. Most authors have interpreted the rhythmites as varvites (Blondeau and Lowe, 1972; Schwab, 1976). However, these rhythmites are somewhat unusual for varvites, having overly thick couplets (up to 150 mm), a thinner "summer" layer than a "winter" layer, and an upward decrease in mean grain size within individul laminae. Thus, Kuenan (1951) suggested that the varves may not be seasonal, but represent deposition of silt by turbidity currents. The alluvial deposits above and below the glacigenic sequences contain poorly sorted, angular, very coarse clasts, pointing to the presence of high stream gradients and mountainous relief. The regionally extensive outcrops of dropstone-bearing rhythmites suggest a network of alpine glaciers flowing down from mountainous terrain to low-lying topography (Schwab, 1976).

10. Southeastern Newfoundland

The Gaskiers Formation (Williams and King, 1975) is found on the Avalon Peninsula (Fig. 10-1) where it has its best exposure and greatest development on the east side of St. Mary's Bay, its type area (Williams and King, 1979). The Avalon Peninsula is part of a much larger zone consisting of north-northeast trending belts of upper Proterozoic volcanic and sedimentary rocks, overlain by lower Paleozoic strata in areas. As with most rocks of the peninsula, the tillites of the Gaskiers Formation have been folded, faulted and cleaved, but only slight metamorphism has taken place. The formation is part of the Conception Group, a 4000 m succession of Precambrian volcanigenic sediments, mostly deep-water turbidites with tuff pillow lavas and rhyolitic flows. The Gaskiers Formation is situated between two tuffaceous siltstone and sandstone units, the underlying Mall Bay Formation and the overlying Drook Formation. The age of the tillites is still disputed, but is estimated to be at least 800 Ma old (Anderson, 1978).

In the type area, the Gaskiers Formation is developed as a fairly continuous series of massive tillites with a few thin intervening beds of sandstone, conglomerate, and laminated siltstones with dropstones (Fig. 10-2). The top of the formation is defined by a red tillite overlain by a thin red mudstone (William and King, 1979); this has not been seen outside the type area, however. The tillites units within the formation are thick (from 3 m to 55 m), massive and structureless, with sharp, conformable lower

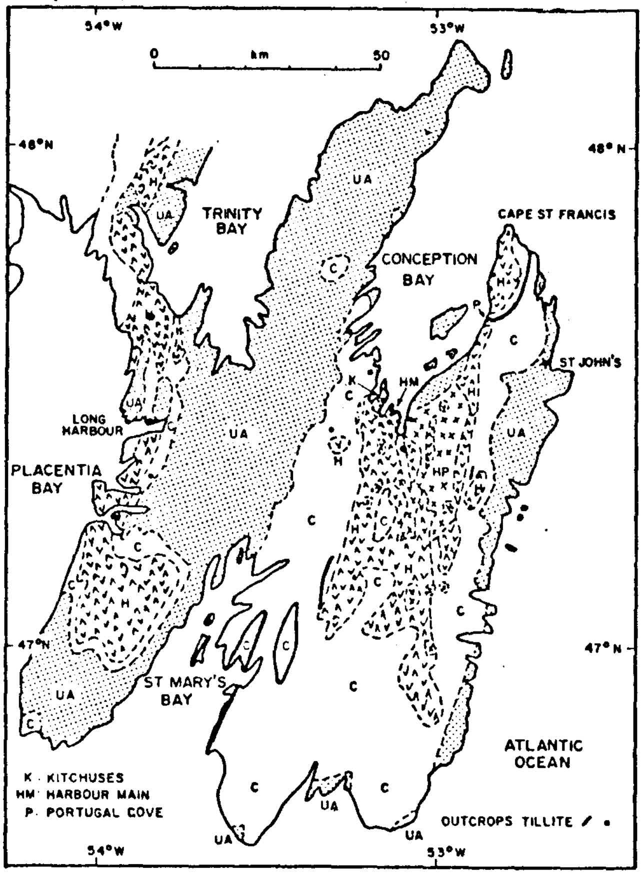

Fig 10-1 Sketch map of the Avalon Peninsula showing the distribution of deposits of the Caskiers Formation. C, Conception Group (marine sediments), H, Harbour Main Group (mainly volcanics), HP, Holyrood Plutonic Series (mainly granite), UA, Upper Assemblage (late Precambrian marine and continental deposits, plus, locally, relics of Late Paleozoic strata). (Williams and King, 1981).

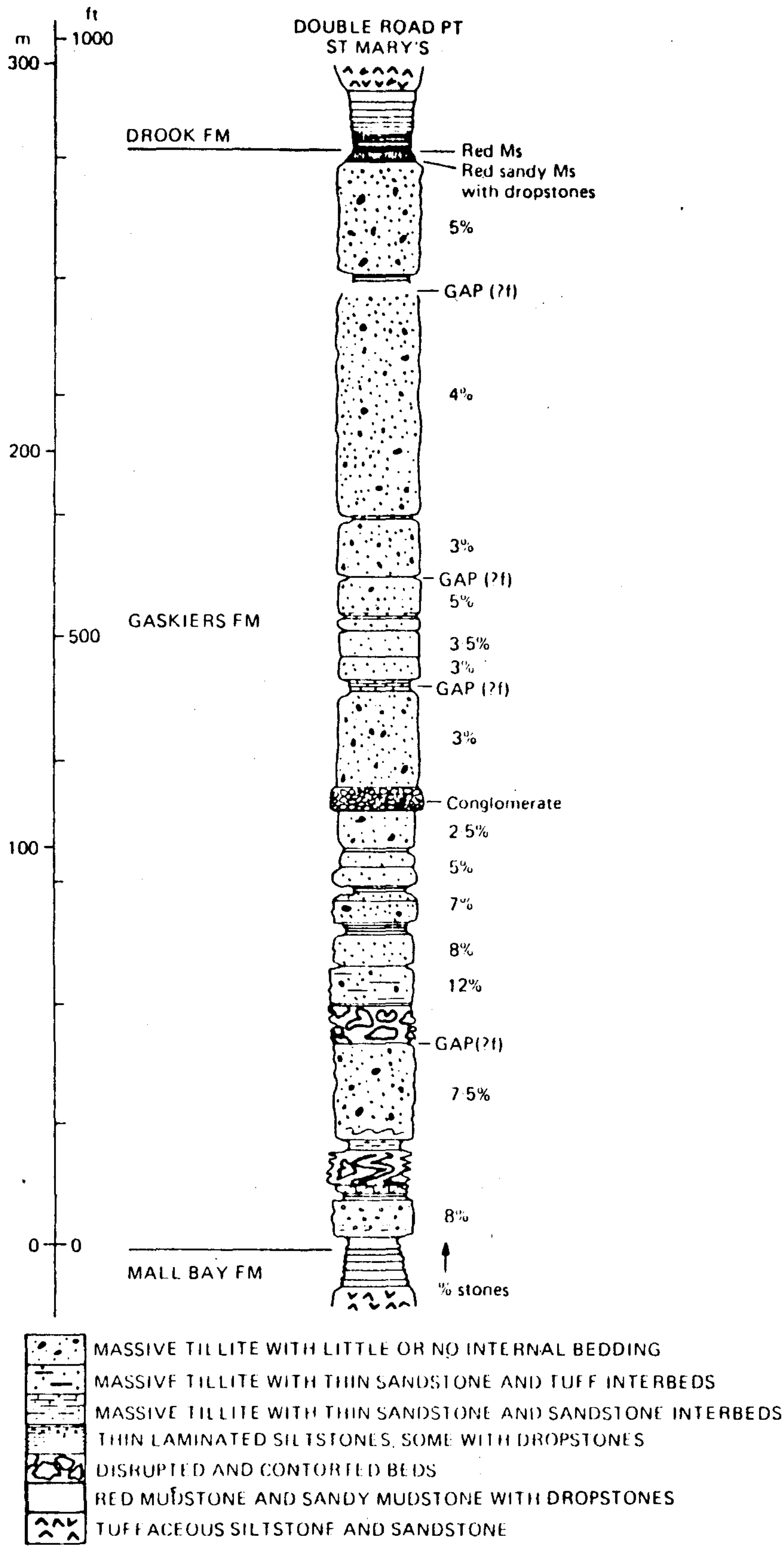

Fig. 10-2 Stratigraphic column of the Gaskieers Formation, taken at Double Point Road, St. Mary's Bay (William and King, 1979).

contacts in most cases. The presence of thin interlayers of sandstone and tuff impart internal bedding in some of the tillite units. The section at Great Colinet Island in St. Mary's Bay shows evidence of volcanic activity during deposition of the tillites, with a layer of agglomerate present in the lower part of the succession, and volcanic debris incorporated into the glacial deposits. Separating the massive tillites are thin, rhythmically bedded, gray to green mudstone, siltstone and sandstone. These rhythmites are usually developed as alternating layers of dark-colored silt to coarse sand and light-colored mud separated by sharp boundaries. Thicknesses of individual layers range from less than a millimeter to several centimeters or more. Dropstones present in the rhythmites have deformed and disrupted layers into which they have fallen. Sandstone beds present between tillite units may be structureless or display convoluted lamination, ripple-drift lamination or slump folding. Direction of sand transport was apparently towards the south to southeast. Stones present within the tillites were mostly igneous and sedimentary types derived from within the basin, but some exotic metamorphic stones were also present. Stones varied in size from pebbles to boulders, but most were less than 10 cm in diameter. In the type area, stones are generally subrounded to well-rounded, and faceted, striated or flat-iron shaped.

The Gaskiers Formation, conformably bounded by the marine sequence of the Conception Group, was itself deposited in a marine environment, and has been interpreted as a sequence of submarine tillites and proglacial sediments laid down by a wet-based ice sheet similar to that of the Ross Ice Shelf, Antarctica (Chriss and Frakes, 1972). Massive tillites and bedded sediments were produced by meltwater turbidity currents with dropstones and till pellets being added by melting icebergs. Other features accounted for by this explanation include the presence of interbeds in some massive tillites, debris flow deposits, lag deposits and load induced folds. Penecontemporaneous volcanism in the St. Mary's Bay area produced pyroclastic material which was then deposited on the surface of the ice sheets or directly into the sea, and was incorporated into the tillite with the rest of the debris. In late Proterozoic times, this area was a broad, shallow sea with a chain of volcanic islands which may have served as centers from which the ice sheets advanced. Alternately, the presence of exotic metamorphic stones may suggest neighboring land areas served as the main ice centers. In any case, the Gaskiers Formation appears to be the result of advances and retreats of a single glacial episode (Anderson and King, 1981).

11. Southern Labrador and Eastern Quebec

The Archaean Grenville Gneiss below the Cambrian Bradore sandstones of the Labrador "series" of southern Labrador and eastern Quebec exhibits a prominently grooved, polished surface believed to be the product of glacial erosion. The grooves on fresh, unweathered gneiss underlie Cambrian sedimentary formations and most predated their deposition. The Cambrian sandstones, limestones and shales rest upon this erosional surface along the south coast of Labrador and Quebec for about 50 km. The grooved surface is carved into Grenville age granites and granitoid gneisses which are generally foliated, folded, coarsely crystaline metamorphics. The sub-Cambrian surface possesses undulating grooved relief up to 10 m (Swett and Smit, 1972), although no glacial stria are observed. The overlying arkosic conglomerates and sandstones contain no evidence of glacial transport, such as faceted or striated pebbles, but were probably deposited in a tidal marine environment, as cross-bedding type and trace fossils suggest.

The probable glacial erosional surface may correlate with glacial deposits beneath similar sandstones in eastern Greenland (Kloftelv Sandstone) and northern Spitsbergen (Tokammane Sandstone). Another possible correlation would be with the erosional surface beneath the Eriball Sandstones in northwest Scotland, but no glacial features have been noted there (Swett, 1981).

12. Eastern Greenland

A group of tillites overlying the Eleonore Bay Group is found in the Kong Oscars Fjord and Kejser Franz Josefs Fjord region in eastern Greenland (Fig. 12-1). This has been placed in a variety of different formations, and was most recently described by Haller (1971) as the Tillite Group, which has become the most widely accepted name. Lying within the Caledonian fold belt, this unit has been subjected to major folding of a simple style and also a great deal of faulting. However, the outcrops lie outside the range of regional metamorphism. An uppermost Precambrian or Eocambrian age has been set for the tillites on the basis of the stratigraphic position between the late Precambrian Eleonore Bay Group and lower Cambrian sediments.

Five divisions are recognized within the Tillite Group. A basal gray massive tillite up to 150 m thick overlies the limestones and dolomites of the Eleonore Bay Group with slight disconformity. Above this are cross-bedded sandstones and grayish-black sandy shales 150 m thick, followed by an upper tillite up to 200 m thick. This second tillite is developed as three or four tillite levels with interbedded sandstones, and is characterized by a hematite-rich matrix in some localities and a predominance of crystalline boulders. Over this are the mudstones and varve-like shales of the Lower Canyon Formation. Higher in this section are black silty shales, giving way to dolomites and limestones with algal structures near the top of the formation. The Spiral Creek Formation, ranging from zero to 55 m thick, completes the Tillite Group. This uppermost unit consists of red mudstones, cross-bedded sandstones with

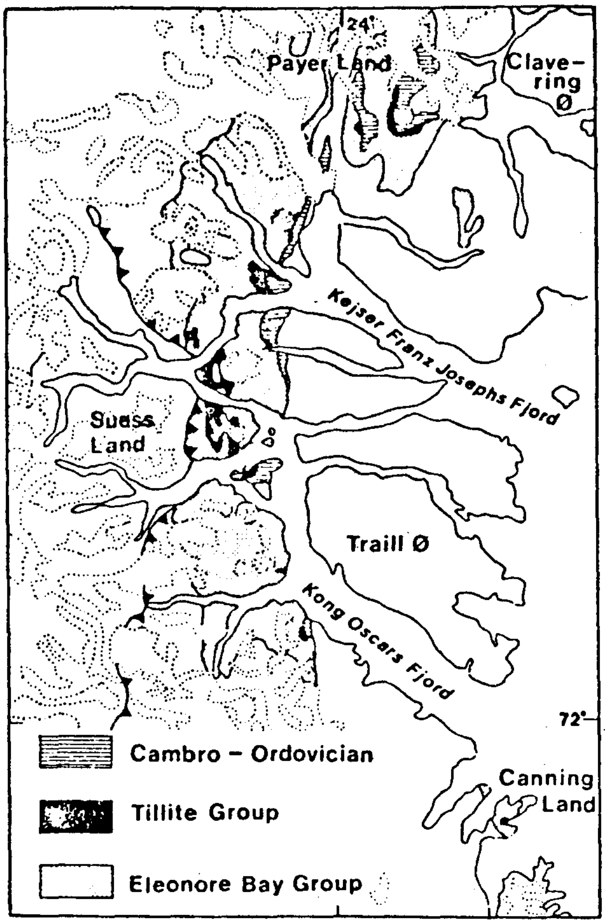

Fig. 12-1 Distribution of the Tillie Group in East Greenland. Glaciers and ice cap boundaries are shown by dotted lines. (Higgins, 1981).

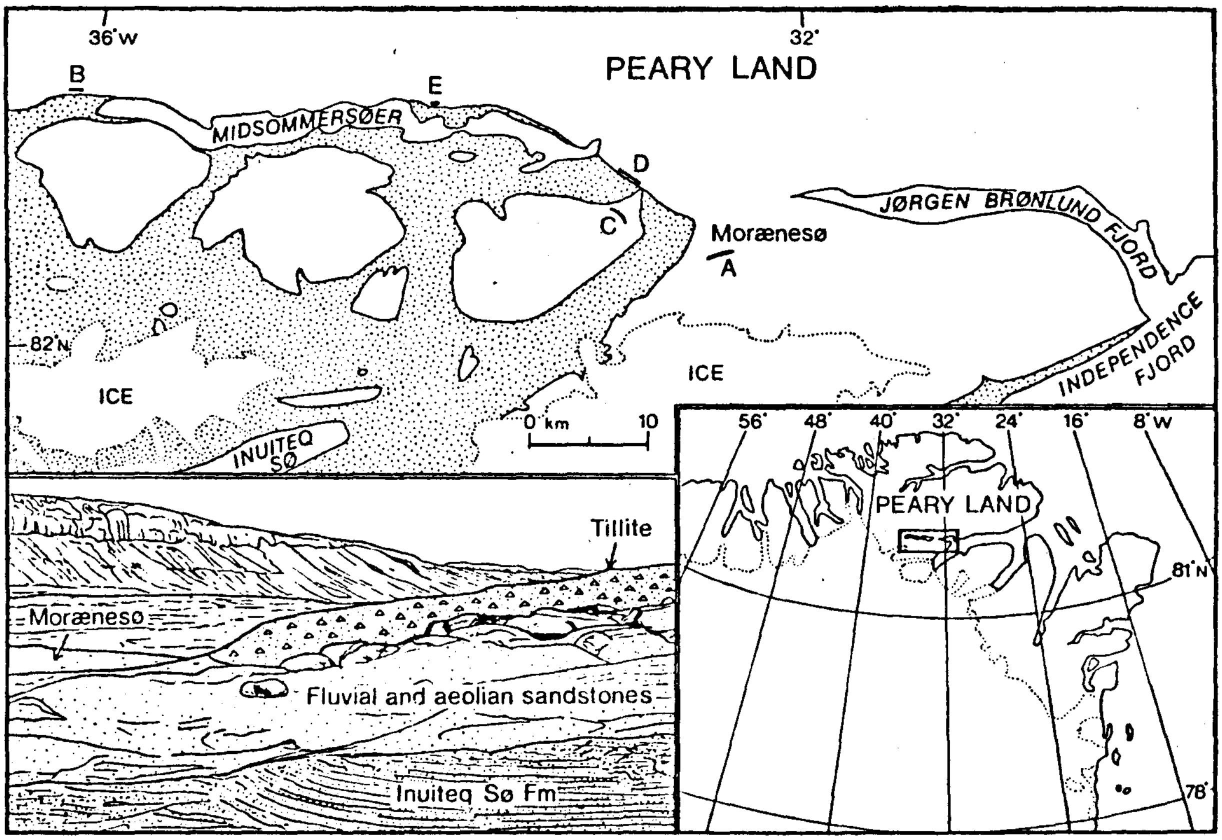

Fig. 13-1 Distribution of supposed tillites of the Moraeneso Formation (black and labeled A - E) and older rocks (stippled). Left inset shows field appearance of the tilloid of the type locality (A). Right inset shows location in North Greenland. (Clemmensen, 1981).

mudcracks, and a gypsum-bearing limestone breccia. Separated by a slight unconformity, the basal unit of the lower Cambrian succession overlies the Tillite Group. Both the lower and upper tillites are largely massive, unstratified and resistant, with minor sorting and sandy or conglomeratic lenses. Ripple-marks and cross-bedding are observed in all but the lower tillite, and some varve-like laminations are apparent in the lower Canyon Formation. Stones within the lower tillite are largely traceable to local dolomite and limestones sources, while stones in the upper tillite are traceable to quartzites and granites from progressively lower units of the Eleonore Bay Group. Striated, polished and flat-iron shaped pebbles have been reported.

The Tillite Group overlies the carbonate deposits of the uppermost Eleonore Bay Group, which were laid down in a warm subtidal marine environment. The two separate tillites suggest two glacial episodes, with the intertillite beds being deposited in a shallow sea during the interglacial period. Haller (1971) suggested that the tillites were deposited in a shallow sea from icebergs or glaciers, due to the presence of shallow marine interbeds. Huber (1950) attributed the upper tillite to ground morainic deposits in desert conditions, because of the presence of polished sand grains and 'dreikaufer' pebbles. The base of the Canyon Formation contains varve-like laminations of possible glacial origin, but the limestones and dolomites higher in the sequence suggest warm water conditions (Higgins, 1981).

13. Northern Greenland

The Moraeneso Formation of Jepsen (1971) is named after a lake near its type locality in southern Peary Land, northern Greenland (Fig. 13-1). This tilloid-bearing formation makes up part of the Proterozoic - late Paleozoic platform cover in northern Greenland (Dawes, 1976). Although at times fractured by faulting, the tillites have not undergone significant deformation or metamorphism. The formation overlies the Precambrian Inuiteq So Formation, and underlies the Cambrian or late Precambrian Portfjeld Formation. Cambrian fossils are found about 500 m above the top of the Moraeneso Formation (Jepsen, 1971).

The Moraeneso Formation is divisible into a basal continental sequence with tilloids and/or stratified conglomerates, and an overlying series of shallow marine rocks with local conglomerates (Clemmensen, 1979). The tilloids outcrop in a belt stretching 60 km west of the type locality. As described by Clemmensen in 1978, the tilloids are dark red-brown and massive, with stratified sandstone bodies or irregular sandstone lenses and generally sharp contacts with adjacent facies units. In addition to the tilloids which comprise most of the formation are till-like conglomerates, sandstones with trough-crossbedding and scattered angular clasts, red siltstones, and laminated sandstone-mudstones with dropstones. Above the uppermost tilloid is a thin unit (2.5 to 3 m) of yellow dolostone with stromatolite domes. The tilloids have scattered horizontal laminations, generally observed as light gray sandy layers several centimeters thick. Interbedded sandstones mostly show soft sediment deformational structure, but may contain graded-bedding or trough cross-bedding. Aeolian, fluvial and lacustrine deposition is indicated in rocks associated with the tilloids. Sandstone-siltstone couplets with outsize dropstones point to lacustrine deposits. Aeolian rocks occur as sandstone polygons. Stones present in the tilliods are mostly angular or subangular in shape, and quartzite sandstone or dolerite in composition. Most of the stones appear to have been locally derived, except for a small percentage of granite or gneiss erratics. Pentagonal flat-iron shapes are common amongst stones, but subparallel striations and grooves are rare (Clemmensen, 1981).

The Moraeneso Formation consists of massive tilloid, possibly representing lodgement till, and laminated tilloid, which may be flow till or lodgement till. The lack of sorting, clast shape and fabric, and presence of striated stones all point to a glacial origin for the tilloid, but a non-glacial mode of deposition cannot be completely ruled out. Erosion by fluvial, aeolian and possibly glacial agents took place prior to deposition of the formation, creating hills and depressions over which continental sediments, including tilloids would be laid. Grounded ice sheets produced the tilloids, while non-glacial processes created the other units in the formation. The apparent direction of glacial transport, as determined by till fabric studies, is from the south or southeast. (Clemmensen, 1981).

EXTENT OF LATE PROTEROZOIC GLACIATION

The ice ages of the Upper Proterozoic were far more extensive than those of earlier times, with evidence of glaciation present on every major continent except Antarctica. Close examination of these Upper Precambrian deposits on a world-wide basis reveals them to be of at least three distinct age ranges. By far, the greatest number of diamictites have been dated around 600 Ma; some examples of these are found in West Africa, South Africa, Kazakhstan, South Australia, Scotland, Ireland, Norway, and western U.S.S.R. These represent an epoch of glaciation known as the Varangian ice age, named after the many tillite occurrences near the Varangian Fjord (Kulling, 1951). Another group of glacigenic rocks cluster around the age 750 to 800 Ma, with examples being found in the Changan Formation and the Nantuo tillites of China, the Damara Supergroup of Namibia, and the Sturtian tillites of southern Australia. The earlier event was referred to by some authors as the Sturtian ice age (Dunn, 1971; Harlan and and

Herod, 1974). Still another group of tillites ranges from 900 to 950 Ma. This includes the Varianto diamictites of Namibia and the Kundelunau Supergroup of Zaire and Shaba. Less is known of the extent and duration of the pre-Varangian ice ages.

On the North American continent, most of the well-dated deposits have been placed near the Precambrian/Cambrian boundary. These include the Upper Tindir Group of Alaska, the Mount Lloyd George diamictites of British Columbia, and the Tillite Group of eastern Greenland, and probably some of the more poorly dated deposits as well. An older collection of tillites whose ages range between 800 and 850 Ma include the Toby Formation, the Scout Mountain diamictites, and the Mineral Fork Tillites of western North America, and the Gaskiers diamictites, Grandfather Mountain Formation and the Mount Rogers Formation of the eastern craton. Earlier still, the Kingston Peak Formation of Death Valley was deposited by glacial activity between 900 and 1300 Ma ago. Thus, it appears that the diamictites represent at least three periods of glacial activity. The ages of the remaining tillites mentioned in the report are only known to within 200 Ma or more; thus, the glacial event recorded by them is not known.

The relationship between the Precambrian tillites of North America and those of the rest of the world is still poorly understood for several reasons. The absence of guide fossils makes precise dating and correlation of glacigenic beds extremely difficult. Precambrian paleogeographic studies themselves, whether they involve plate tectonics, patterns of glaciation, or whatever, have been severely hampered by lack of fossils and deformation of rocks. While local or regional correlations can be made with some confidence, worldwide correlations are extremely doubtful. Conflicting data or plate positions and paleolatitudes have created a controversy on the extent and patterns of Varangian glaciation. Bigwood and Harland (1961) found low paleolatitudes for the tillites of eastern Greenland. Similar reports in other deposits plus their extremely widespread occurrences led to speculation of glaciation on an unprecedented scale in latest Proterozoic times, extending even into tropical latitudes. This would be of great importance stratigraphically, since all Precambrian glacial deposits would be the same age and could serve as marker beds. However, McElhinny et al (1974) proposed an alternate hypothesis, that the diamictites reflect a 90o shift in polar position over a period of some 300 Ma. As the pole moved from north to south across South America, a series of glacigenic deposits of decreasing age were laid down in west Africa. Briden and Gass (1974), however, present evidence for a polar wandering path from south to north for the same period. It is obvious, then, that considerably more paleomagnetic data and precise age dates from the glacial sediments themselves are needed to clear up this controversy.

EARLY PROTEROZOIC GLACIATION IN NORTH AMERICA

The earliest evidence of widespread glaciation on the North American continent is found in deposits dating between 2500 and 1700 Ma old. The Gowganda Formation of the Lake Huron region contains the best-known tillites of the glacial event. The well-preserved sedimentary structures indicative of glacial deposition, such as laminated argillites and dropstones, striated stones and pavements have made this unit an object of particular interest. Other early Proterozoic diamictites are found in the Hudson Bay area of the Northwest Territories, in southeastern Wyoming, northern Michigan, southern Ontario, and northern Quebec (Fig. B). Diamictites of the same general age also occur in the Black Hills of South Dakota (Kurtz, 1981). However, there is little evidence to strongly indicate a glacigenic origin for these deposits; hence, they will not be included in the discussion which follows.

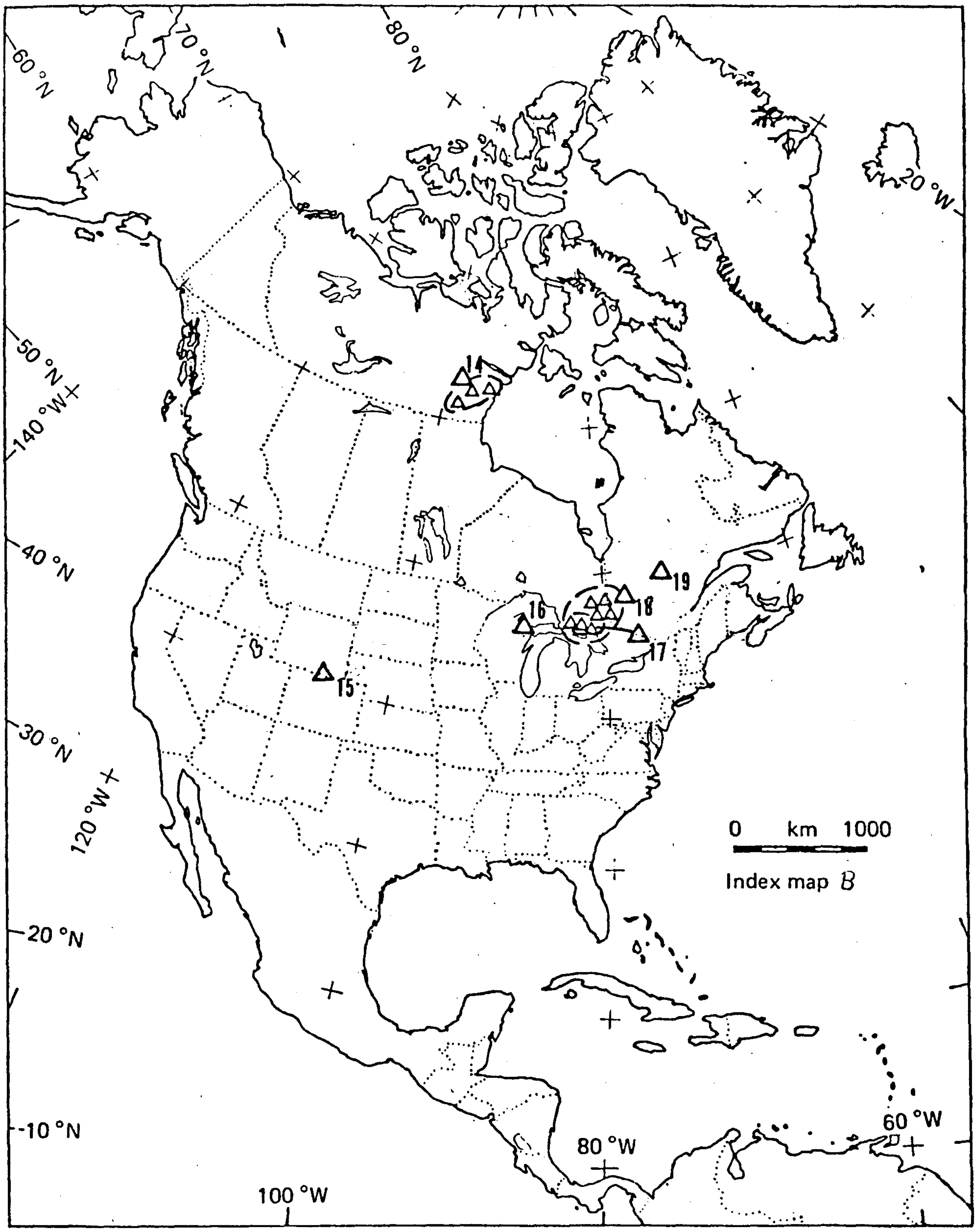

Fig. B Distribution of early Proterozoic glacigenic rocks of uncertain age in North America. Numbers refer to sections of text. (Modified from Hambrey and Harland, 1981).

14. Northwest Territories, Canada

Along the western shores of Hudson Bay between latitude 61°N and $62^{\circ}31'$N lie outcrops of the Hurwitz Group, a series of siltstones, sandstones and pebble conglomerates with a basal diamictite unit (Fig. 14-1). This group is preserved in downwarps with moderate to steep dips. Although the unit has been subjected to variable metamorphic processes during the Hudsonian orogeny, stratigraphic integrity is preserved and primary sedimentary structures are still observable. The lowermost unit, the Padlei Formation, was formed by glacial processes between 1810 Ma and 2550 Ma, as determined from geochronometric dates of surrounding igneous rocks (Wanless and Eade, 1975).

The Padlei Formation rests unconformably on Archaean volcanic, metasedimentary, and plutonic rocks in some places, and on quartz arenite, siltstones and polymictite conglomerates of lower Proterozoic age elsewhere. The entire formation is about 500 m thick and usually consists of a lower massive diamictite member from zero to 400 m thick and an upper finely-bedded and laminated member up to 200 m thick (Fig. 14-2) (Bell, 1968). The lower member is a gray and green-gray sandy pebble-, cobble- and boulder-bearing diamictite with minor interbeds of siltstone, sandstone and conglomerate. The upper member is largely sandstones, siltstones and argillites with rhythmic laminations and isolated clasts of basement rock types. The

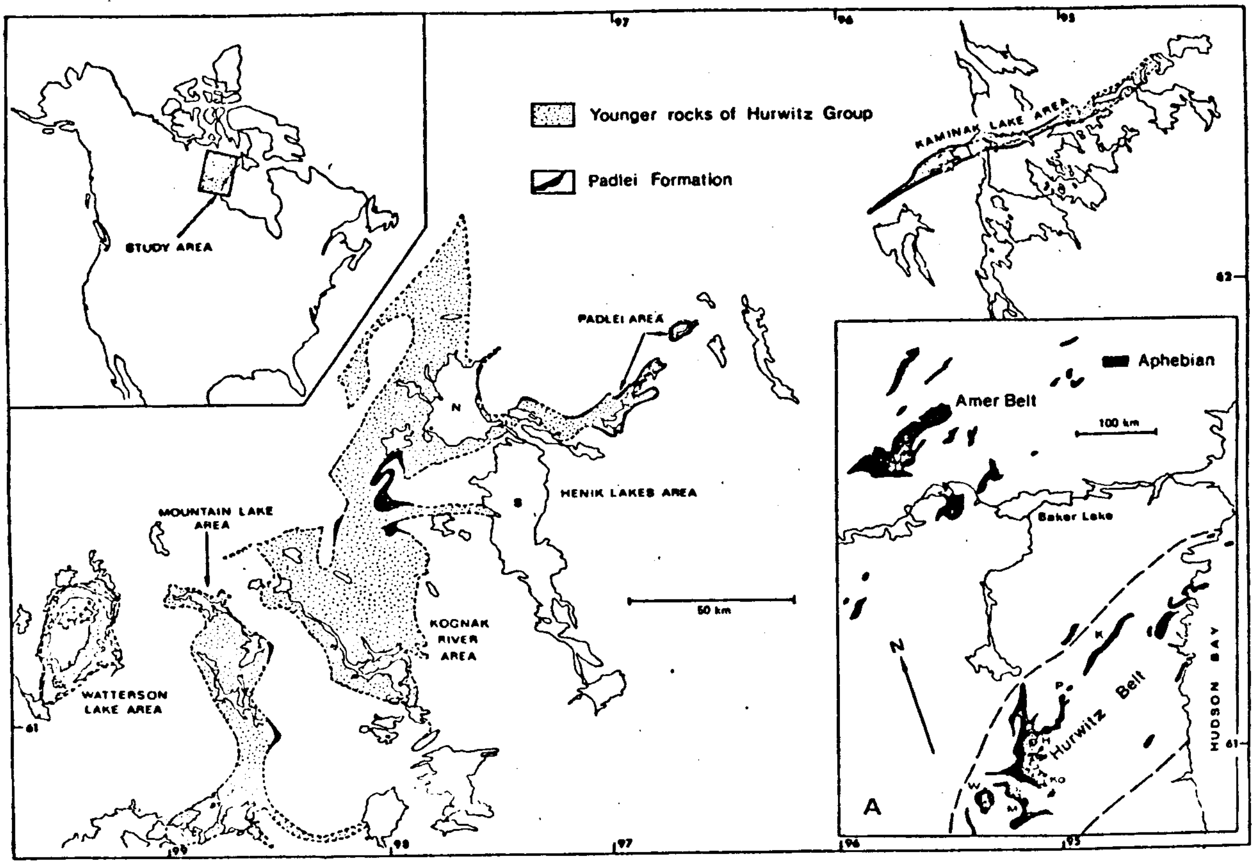

Fig. 14-1 Sketch map showing the distribution of the Padlei Formation (black) and younger rocks of the Hurwitz Group (stippled). Inset A shows the location of the Hurwitz Belt (Bell, 1970).

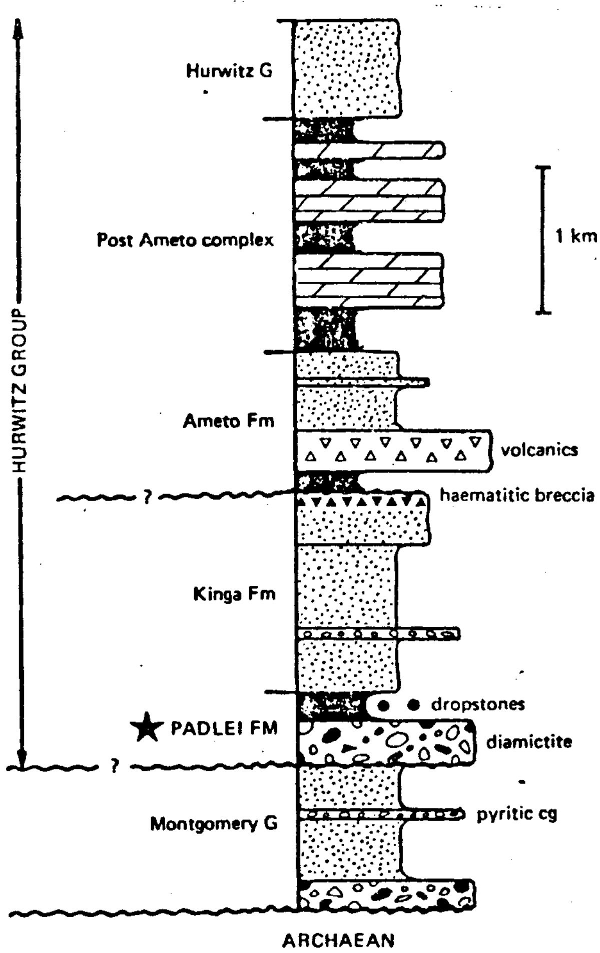

Fig. 14-2 Schematic representation of the Hurwitz and Montgomery Groups. Star indicates position of the Padlei Formation. Note diamictites in the Montgomery Group, and hematitic breccias at the top of the Kinga Formation. cg, conglomerate; Fm, formation; G, group. (Young et al, 1981).

limited geochemical work performed on the Padlei Formation (Young, 1973) points to the minerological immaturity of the rocks and has been compared with the results of the Gowganda Formation (McLennan, Fryer and Young, 1981).

A glacial origin for the Padlei diamictites was proposed by Young (1973). The discontinuous bodies of diamictite with variable clast composition might represent deposition from different glacial lobes. The general lack of bedding in the diamictites suggests deposition in a terrestrial environment, possibly at the edge of a continent al ice sheet. The general paleoslope, as deduced from paleocurrent measurements in associated rocks, was to the northwest (Young, 1973; Young and McLennan, 1981). The upper member of the formation was most likely deposited in a shallow marine basin or in glacial lakes that formed after the melting of the ice sheet responsible for the deposition of the diamictite. The isolated clasts have been interpreted as dropstones, suggesting the persistence of the ice sheets during the deposition of the upper member of the Padlei Formation (Young and McLennan, 1981).

15. Southern Wyoming

Outcrops of Precambrian rocks, including lower Proterozoic metasediments and diamictites, occur in the Sierra Madre and Medicine Bow Mountains in Southeastern Wyoming (Fig. 15-1). Diamictite-bearing units include the

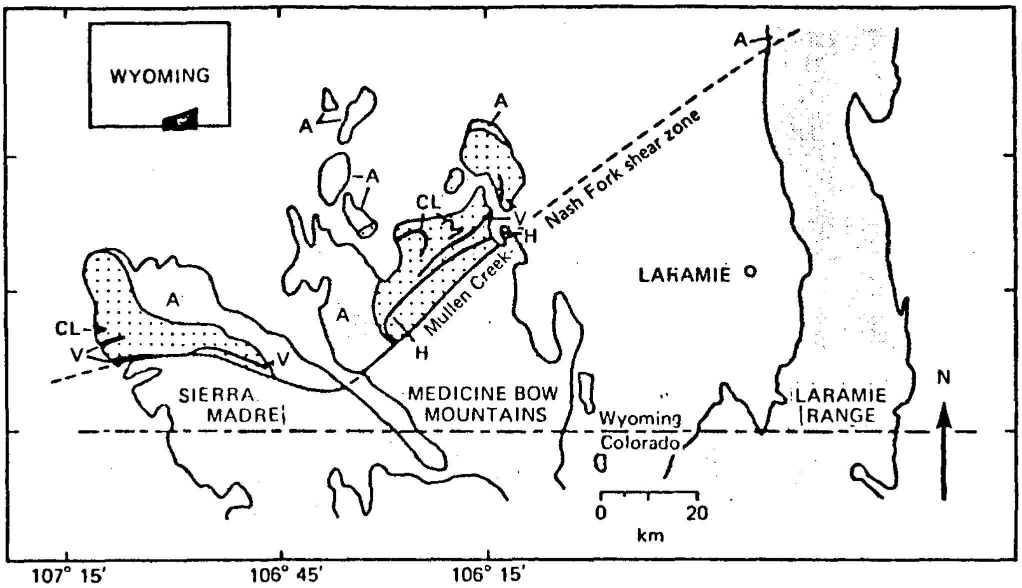

Fig. 15-1 Location of Precambrian rocks (shaded area) and Proterozoic metasediments (stippled area) in southern Wyoming. Diamictite-bearing formations marked in black. A, Archaean basement; CL, Campbell Lake Fm.; V, Vagner Fm.; H, Headquarters Fm. (houston, et al, 1981).

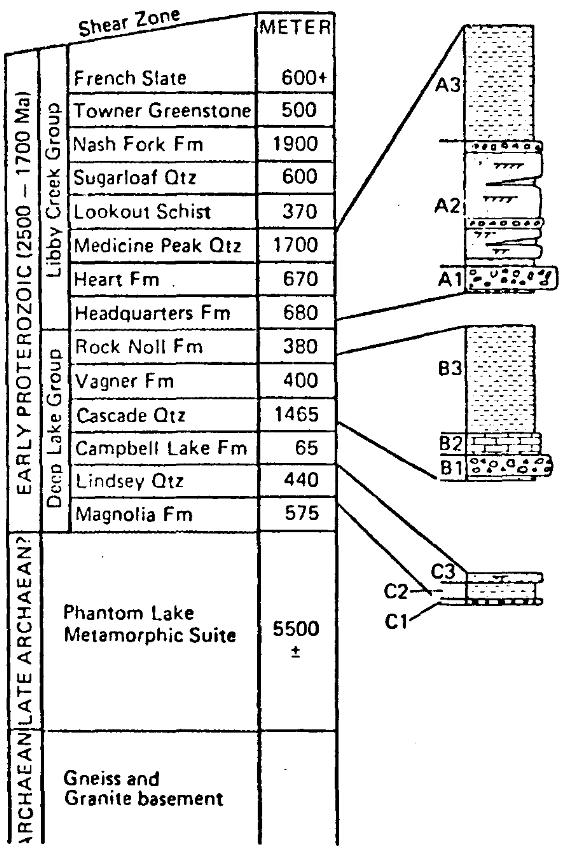

Fig. 15-2 Outline of the stratigraphy of the early Proterozoic metasediments of southern Wyoming showing stratigraphic setting of the diamictites. Diamictites identified by pattern of clasts, quartzite by stipple, laminated phyllite by dashes, marble by block pattern (Houston et al, 1981).

Campbell Lake and Vagner Formations of the Deep Lake Group (Karlstrom and Houston, 1981) and the Headquarters Formation (Blackwelder, 1926) of the Libby Creek Group. The diamictites are preserved in troughs or synclines north of the Mullen Creek - Nash Fork shear zone, and occur within miogeosynclinal sediments resting on an Archaean granitic gneiss basement. This miogeosynclinal succession has been metamorphosed and deformed extensively, but to a lesser degree in the north central Medicine Bow Mountains. Diamictites of the Deep Lake Group are discontinuously exposed from the western Sierra Madre to the eastern central Medicine Bow Mountains, while those of the Headquarters Formation crop out in the central Medicine Bow range only. The stratigraphic relationship of the formations is shown in Fig. 15-2. (Houston et al, 1981).

As seen at their best exposure in the Medicine Bow Mountains, the diamictites form the basal units of each of the three formations, and are bounded by a basal unconformity in each case. The oldest diamictite, that of the Campbell Lake Formation, is an unstratified or poorly stratified arkosic quartzite with clasts of granite, phyllite and quartzite. This is overlain by phyllite and phyllite quartzite with local cross-bedding. In the Vagner Formation, the diamictite is subarkosic with angular clasts of granite, quartzite and schist. It has a laminated base with scattered stones (dropstones?) and grades upward into a massive to very poorly stratified paraconglomerate. This is conformably overlain by a chlorite-biotite phyllite member. In the Headquarters Formation, the diamictite is developed as finely-laminated chlorite-biotite-quartzite phyllite, with a laminated basal portion grading upward into a massive paraconglomerate much as in the Vagner Formation. A medium-grain arkosic quartzite member with diamictite and phyllite interbeds rests conformably upon this unit, and is overlain in turn by a laminated biotite-chlorite quartz phyllite member. The stones present in all of the diamictites are subrounded to angular, typically granite, phyllite and quartz with metabasalts and basalts in the Campbell Lake and Vagner Formations, respectively. Although quartzite underlies the diamictite in all localities, the stones are mostly granitic except in the upper two diamictites of the northeastern Medicine Bow Mountains (Houston et al, 1981). Nevertheless, it appears that the stones were derived locally from older rocks of the area. Possible striated or "soled" clasts have been reported (Blackwelder, 1926) but not confirmed.

The origin of the diamictites of the Campbell Lake Formation is difficult to ascertain, but the presence of dropstones and phyllites suggests glacial deposition. Carey and Ahmad (1961) proposed that the diamictite, marble and phyllite succession of the Vagner Formation was produced by deposition from a dry-based glacier, with the diamictite representing till deposited in a floating ice shelf and iceberg zone, while the marble was attributed to carbonate brine deposits in an iceberg zone. In the Headquarters Formation, the diamictites appear to be glaciomarine in origin, but may also be till deposited by wet-based glaciers. The quartzite member of the Headquarters Formation could have been laid down by the advancing and retreating wet-based glacier, as till and fluvioglacial deposits are also included in this unit. The overlying phyllite member is believed to have been deposited seaward of the area of glacial influence, as it grades upward into the deltaic quartzites of the Heart Formation. Lithologically, the diamictites strongly resemble the Huronian succession of the north shore of Lake Huron (Young, 1970), and have been dated at roughly the same age (2100 to 2500 Ma).

16. Northern Michigan

Lower Proterozoic glacial deposits occur in the Fern Creek, Beany Creek, and Enchantment Lake Formations of northern Michigan (Fig. 16-1). The Fern Creek Formation (Pettijohn, 1943) crops out along the flank of the Menominee iron range east of the city Iron Mountain. The Beany Creek Formation (Puffett, 1969) occurs in the Dead River Basin area, north of the Marquette synclinorium. The Enchantment Lake Formation (Gair and Thaden, 1968; Gair, 1975) is found in the Marquette synclinorium. These glacigenic deposits are preserved in westward trending synclinal troughs or

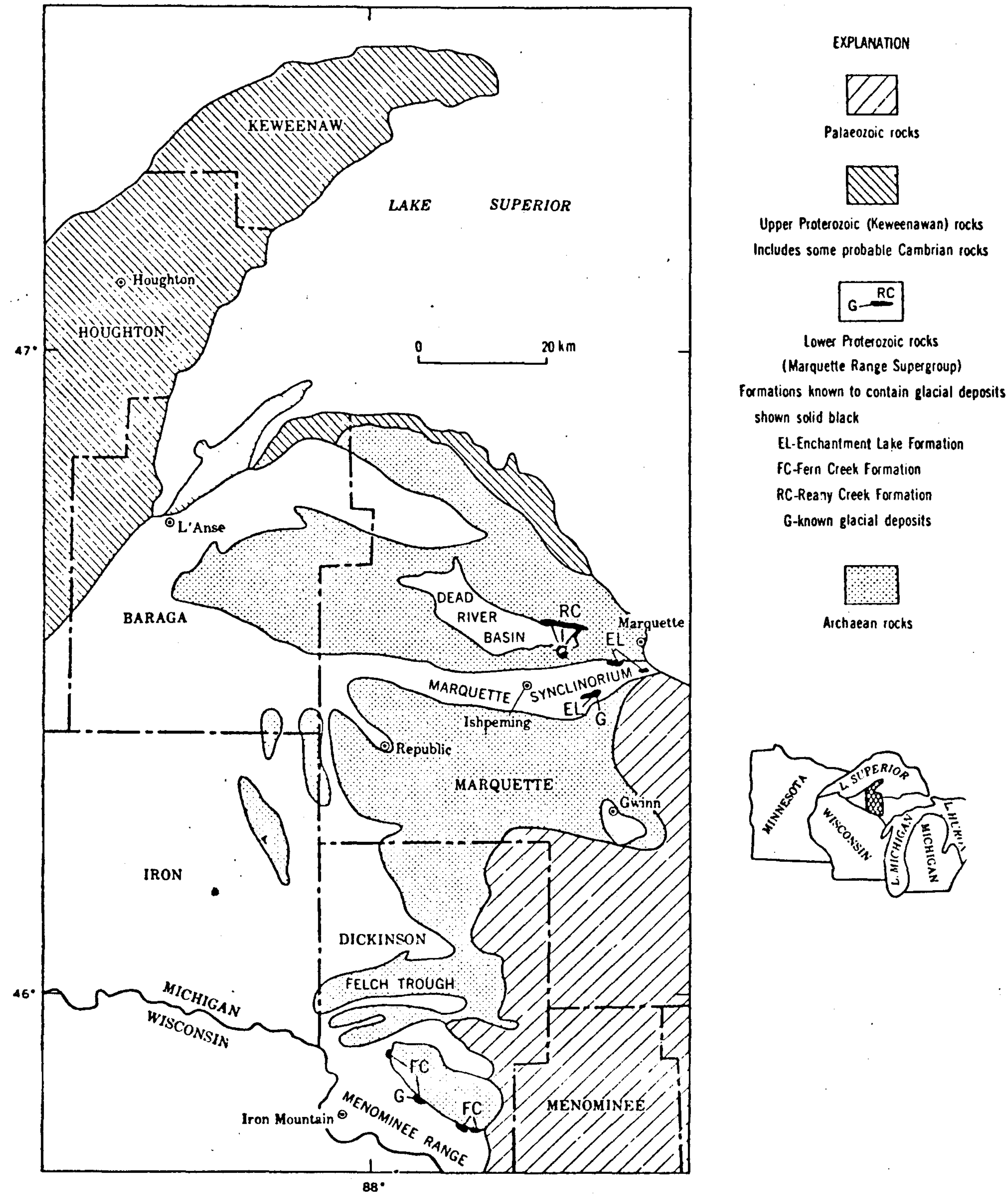

Fig.16-1 Distribution of lower Proterozoic formations containing glacial deposits in the western Upper Peninsula of Michigan (Gair, 1981).

trough segments surrounded by Archaean basement rock. The formations exhibit slight chlorite-grade metamorphism and slight to strong schistosity. Both the Fern Creek and Enchantment Lake Formations are lenticular, the latter consisting of intertonguing beds of various lithologies. These lower Proterozoic rocks are estimated to be 2000 to 2100 Ma old (Gair, 1981).

The lower half of the Fern Creek Formation is largely conglomeratic, but alternates with non-conglomeratic beds. One bed which clearly indicates the glacial origin of the unit consists of angular to rounded pebbles and cobbles, apparently dropstones, in laminated argillite. Other beds of the unit include massive and laminated greywacke, arkose, sericite schist and quartzite. The formation becomes increasingly quartzitic as it grades into the overlying Sturgeon Quartzite. The Enchantment Lake Formation is conglomeratic in the lower part of the section, with pebbles and cobbles of granitic gneiss and greenstone. The middle portion of the formation consists of greywacke, wacke, arkose and subarkose, while the upper part contains sericite, wacke, slate and quartzite, with the last becoming more dominant upward as the unit grades into the overlying Mesnard Quartzite. In some localities, very coarse boulder conglomerate overlies interbedded quartzite and slate, and laminated conglomeratic slate contains angular gneiss dropstones. The Reany Creek Formation is comprised of a basal conglomerate, a middle slaty unit with minor amounts of greywacke, arkose, and scattered grantitic pebbles and boulders, and an upper portion of interbedded arkose, quartzite, slaty greywacke and conglomerate (Puffett, 1974; Clark et al, 1975). The basal conglomerate contains abundant greenstone and granitic boulders in an arkose matrix. The middle slaty unit is massive except when interlaminated with arkose or greywacke. Some beds are varve-like and contain dropstones, principally granitic pebbles and boulders. All three formations have sharply unconformable basal contacts with Archaean gneiss and greenstone basement. (Gair, 1981).

The isolated and rather small extent of each deposit, together with the wide area of which together they are found, suggests that they are products of local valley glaciers formed during a widespread controlling event such as regional uplift. (Gair, 1981).

17. Southern Ontario, Canad a

The Gowganda Formation (Collins, 1917; Quirk, 1917) is a classic glacigenic succession along the northern shores of Lake Huron. The formation outcrops in three areas, as can be seen in Fig. 17-1, one in the Bruce Mines - Blind River area (Area 1), the Espanola - Whitefish Falls area (Area 2), and the Cobalt - Gowganda area (Area 3). The Gowganda Formation was deposited over Archaean basement found to be 2500 Ma old, and was intruded by sills and diabase dikes

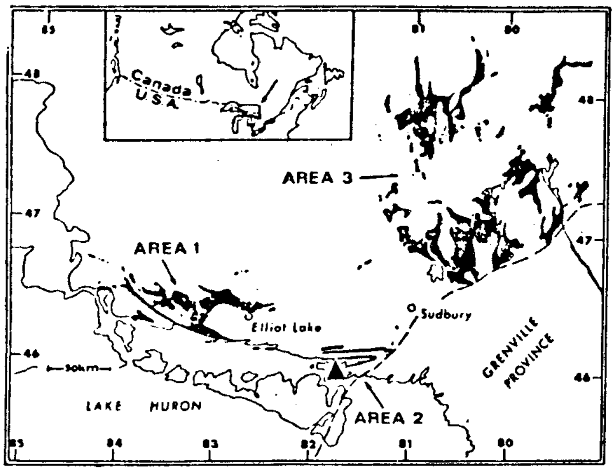

Fig.17-1 Sketch map of the area north of Lake Huron, showing the distribution of the Gowganda Formation (black). Dashed line shows boundary of the Grenville province (Young, 1981).

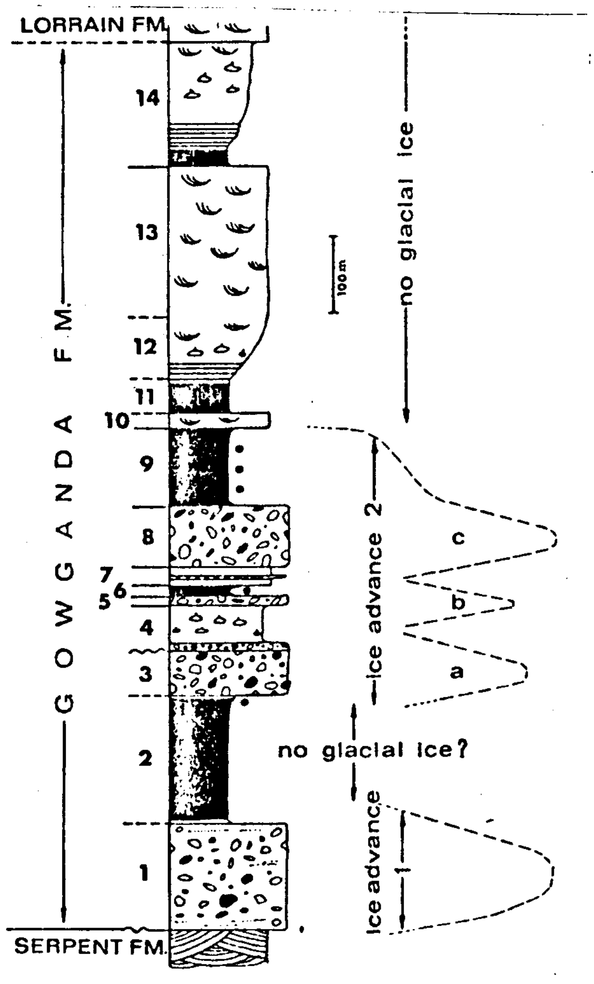

Fig.17-2 Schematic representation of a stratigraphic section through the Gowganda Formation south of Whitefish Falls. Black represents laminated argillites. Width of column corresponds to grain size as follows: widest parts are conglomeratic, intermediate widths represent sandstones and narrowest parts are mudstones. Black dots at right hand side represents dropstones. Types of contacts between units represented on left hand side of column as follows: continuous line, sharp contact; dashed line, gradational contact; wavy line, erosional contact.(Young, 1981).

dated at 2100 Ma (Van Schmus, 1965). To the north, the basal contact of the unit is sharply unconformable against the Archaean basement, while to the south the contact is interbedded with the quartzites of the underlying Serpent Formation. The upper boundary is generally transitional into the sandstones of the overlying Lorrain Formation, but the upper part of the Gowganda has been removed by post-Huronian erosion in the most northerly regions of the Cobalt - Gowganda area. The formation has been mildly metamorphosed to lowest greenschist facies in the northern outcrops, while having been more strongly altered in Area 2 (to amphibolite facies). Nevertheless, sedimentary structures are generally well-preserved.

The stratigraphic succession is highly variable in Areas 1 and 3. Thompson (1966) described a two-fold division in the Cobalt region, consisting of a lower conglomeratic and an upper fine-grained, finely bedded unit. In Area 2 a two-fold division is also apparent, but its relationship to the divisions in the Cobalt area is not clear. A stratigraphic section is shown in Fig. 17-2, based upon exposures in Area 2, south of the village of Whitefish Falls (shown by triangle in Fig. 17-1). The sequence begins with a massive gray diamictite, with boulder-sized clasts and minor stratified sandstone and orthoconglomerate interbeds (Fig. 17-2, Member 1). This is overlain by a laminated to finely bedded gray and green argillite with parallel, wavy and lenticular bedding (Member 2). As can be seen in Fig. 17-2, this argillite unit shows no evidence of glacial origin except for the presence of a few lonestones and diamictite lenses near the top of the member. Member 3 is a gray to blue-green massive diamictite, suggesting a return of glacial conditions. Member 4 is a gray to buff interbedded orthoconglomerate with boulder size clasts, which is followed by several minor members consisting of siltstone, sandstone, diamictite, argillite and conglomerate (Members 5 to 7). Member 8 is a gray-green massive diamictite with pebbles and cobbles, separated from underlying units by a sharp contact. The youngest unit bearing evidence of glacial origin, Member 9, is a finely laminated green argillite with abundant slump structures, contorted bedding and dropstones. The five uppermost members are argillites, siltstones and sandstones with trough crossbedding and ripple cross-laminations, which appear to represent two major deltaic cycles.

A glacial origin for this formation is indicated by the presence of a striated basement, the widespread occurrence of massive polymictic conglomerates with many plutonic rock fragments, the chemical and minerological immaturity of the conglomeratic matrix minerals (Young, 1969), and the presence of finely laminated argillites with common plutonic clasts, without evidence of concurrent volcanic activity to provide a source (Young, 1970). The massive diamictites represent deposits of grounded glaciers with stratified interbeds reflecting floating ice conditions (Members 1 and

3). The interbedded orthoconglomerates of Member 4 were the product of fluvioglacial outwash, representing a minor regression in the ice sheet. Another advance of the ice sheet is shown by the diamictites in Member 5, while a regression is seen in the silty argillites and siltstones of Member 6, laid down subaqueously and bearing dropstones from floating icebergs. Member 7, similar to Member 4 in composition, is another outwash deposit. The massive and stratified nature of Member 8 demonstrates its deposition by both grounded and floating ice. The slumped argillites and dropstones of the uppermost glacigenic unit were probably deposited at a distance from the ice front by floating icebergs. (Young, 1981). An alternate explanation for the diamictites and associated rocks has been proposed by Card et al (1977), who suggest a marine mudflow or debris-flow mechanism. While such mechanisms have been important locally, signific ant relief or tectonic activity at the time of sedimentation would be required. However, the low dips of the outcrops and gradual overlap of the Gowganda Formation onto the basement rocks point to an absence of significant relief at the time of deposition (Dott, 1961). In addition, most of the associated sediments were deposited in shallow water conditions, while subaqueous mudflows are found in deeper water deposits. The wide extent of the diamictites and their stratigraphic position over fluvial or aeolian sandstones of the Serpent Formation led Long (1976) to interpret the diamictites as the result of glaciation.

18. Lake Huron, Ontario, Canada

The Ramsey Lake and the Bruce Formations of the northern shores of Lake Huron (Fig. 18-1) occur in the lower part of the Huronian Supergroup of the early Proterozoic. The Huronian succession was deposited in a partially fault-bounded basin at the southern margin of the Superior Province of the Canadian Shield. These rocks were folded and metamorphosed to varying degrees, with more pronounced deformation in the southern outcrops. The Ramsey Lake and Bruce Formations make up the basal units of the Hough Lake and the Quirk Lake Groups, respectively, which are estimated to be between 2300 and 2500 Ma old. (Young, 1981).

The Ramsey Lake Formation consists of unstratified diamictite from 1 to 150 m in thickness, with a matrix ranging from greywacke to orthoquartzite. Its basal contact is variable, from unconformable against Archaean basement in the north to disconformable or conformable against older Huronian rocks in the southern margins of the outcrop belt. Stratification and cross-stratification is locally present in the upper part of the formation. The Bruce Formation is another pebble- and cobble-bearing diamictite in a poorly sorted, often pyritic matrix of quartzwacke. Evidence of contemporaneous deformation, such as ball and pillow structures and wispy boundaries, may be seen in sandy interbeds in the formation. Basement clasts, apparently dropstones, have been reported in the Bruce Formation. Each

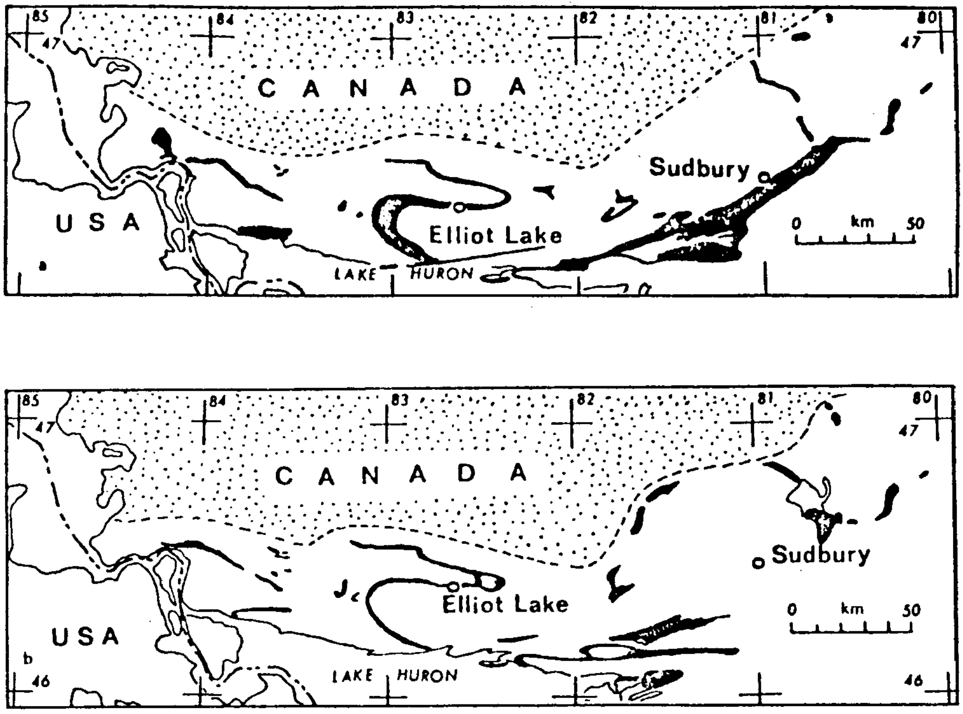

Fig. 18-1 Sketch map of the region north of Lake Huron. a, Distribution of Hough Lake Group, including Ramsay Lake Fm., shown in black. b, Distribution of Quirke Lake Group, including Bruce Fm. Source terrain of Archaean rocks is stippled. (Young, 1981).

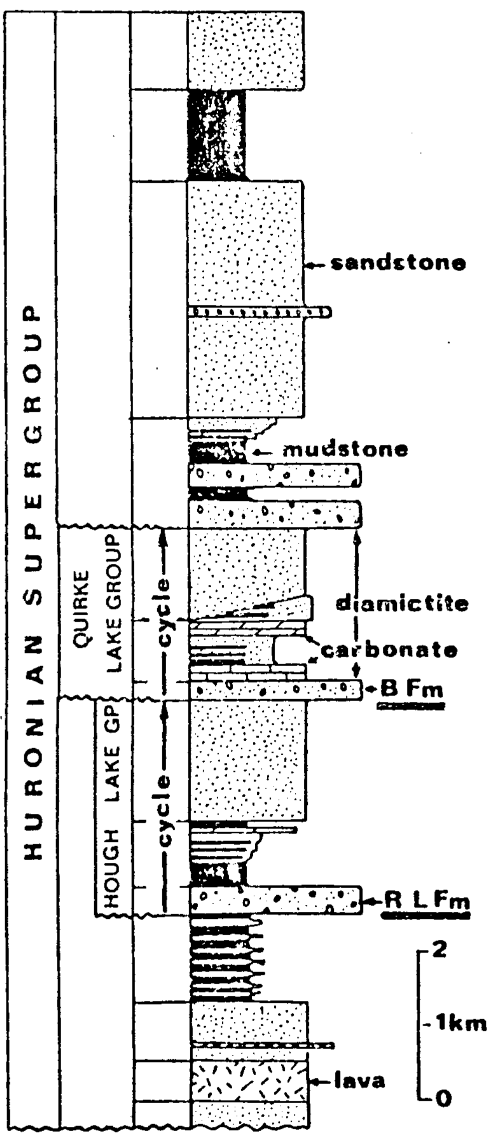

Fig. 18-2 Schematic representation of the Huronian succession showing the stratigraphic position of the Bruce Fm. (B Fm) and the Ramsay Lake Fm (R L Fm) (Young, 1981).

of the formations comprises the basal part of a depositional cycle, as seen in Fig. 18-2. In each instance, the diamictite is succeeded by fine-grained, stratified rocks such as siltstones and mudstones, with carbonates in the Bruce Formation. These are followed by fluvial cross-bedded sandstones, completing the sequence (Young,1981).

Modes of origin cited for the Ramsey Lake and Bruce Formations range from mudflows or debris flow mechanisms to continental and marine glaciation (Farey and Roscoe, 1970). The generally massive nature of the units, their widespread occurrences, and the presence of extra-basinal clasts and dropstones in stratified beds all suggest glacial deposition for the formations. A grounded ice shelf appears to be the most probable depositional agent, but the stratified beds indicate some reworking by water as well. Paleocurrent studies suggest a southerly transportation. (Young, 1968).

19. Northern Quebec

The Chibougamau Formation (Long, 1973) of northern Quebec is a lower Proterozoic clastic sequence exposed in the vicinity of Chibougamau, Quebec (Fig. 19-1). The formation is preserved as scattered remnants of an earlier, more extensive deposit, typically on downthrown sides of major faults. The unit has been subjected to greenschist-grade metamorphism, with epidote-amphibolite metamorphism in areas near the Grenville Front. Situated above Archaean basement

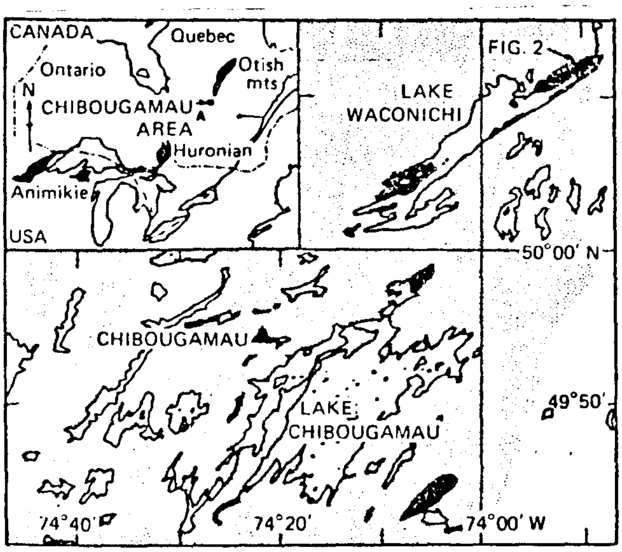

Fig.19-1 Location of rocks of the Chibougamau Formation (black); older rocks (stippled). Inset shows Chibougamau Area in relation to other areas of early Proterozoic rocks (Long, 1981).

rocks dated at 2500 Ma and beneath the Mistassini Group containing the Temiscamie Iron Formation with a minimum age of 1800 Ma, the formation is believed to be of equivalent age to the Gowganda Formation of Ontario (Long, 1981).

The scattered outcrops and abrupt facies changes of the Chibougamau Formation make establishment of regional stratigraphy difficult. In a division proposed by Long (1973, 1974), the formation would be divided into upper and lower members dominated by conglomeratic sandstones and conglomerates, and a middle glacigenic unit of sandstones, diamictites, graded laminates with dropstones, and conglomerates. These divisions are based not on a three-fold sequence present in any one area, but on the lateral equivalence of the glacigenic sequences. Subsurface data from northern Lake Waconicki yielded the best information on the stratigraphic relationships of the beds of the middle member. The sandstones vary from very coarse to very fine, and are intimately associated with intact framework conglomerates. Sandstones are commonly massive to poorly laminated, with little plane and ripple laminations or cross stratification. Diamictites consist of a coarse fraction of granule to boulder size fragments set in a sandy mudstone matrix, with bed thicknesses from a few centimeters to several meters. The diamictites generally are without regular internal structure, and only exhibit preferred orientation of clasts closely associated with large rafts of graded laminite oriented parallel to the bedding. Contacts at the base of the diamictite beds are either sharp or transitional, with the latter formed by incorporation of underlying sediments into the bed. Graded laminites are usually poorly sorted, conglomeratic sandy mudstones with similar size ranges and distributions as the diamictites. The marked stratification and close resemblance to the microvarves of glacial lacustrine sequences allow these to be distinguish readily from the diamictites. The laminites are developed as couplets consisting of a lower coarse silt to coarse sand layer grading into an upper silt or clay layer. Individual couplets range from 2 to 80 mm thick and exhibit considerable lateral continuity. Discontinuous laminae of sandstone and siltstone with ripple cross-lamination are interbedded with the laminites. The clasts within the diamictite units range from granules to boulders, with cobbles and pebbles being the most common size. Pink and white gneissic is the most common rock type of the clasts, the source of which was probably Archaean rocks to the north and northwest. Greenstone clasts are locally abundant, reflecting the bedrock of the area. While larger stones are well-rounded, pebbles and granules tend to be rounded to subangular. Stones are only occasionally facet and never striated, but this could be related to the relative hardness of the rock-types and not an indication of nonglacial deposition. (Long, 1981).

Long (1974) considered the middle member of the Chibougamau Formation to have been deposited at the digitate margin of an extensive, early Proterozoic ice sheet. The presence of numerous dropstones, including some of "till-like material" in the laminites points to their glacigenic origin. The only evidence for glacial deposition of the diamictites are their intimate association with the laminites; the diamictites could represent either tills or laminites remobilized by sediment gravity flows. Deposition by wet-based glaciers allows the best explanation for the origin of the diamictites and graded laminites, while permitting the graded sandstones to have been formed by turbidity currents. The conglomerates of the upper and lower members were interpreted as alluvial fans laid down in a paraglacial environment.

EXTENT OF HURONIAN GLACIATION

The lower Proterozoic series cn the Canadian Shield contains evidence of three distinct episodes of glaciation. The most extensive deposits were left by the latest glacial advance some 2300 Ma ago. While the preserved glacial record is fragmentary, early Proterozoic glacigenic rocks occur over a sizable portion of the craton. Examination of direction of transport shows a radial pattern of movement, suggesting deposition at the margins of a continental ice sheet. Thus, glacial debris apparently accumulated in subsiding basins at the periphery of the Canadian Shield in early Proterozoic times (Young, 1979).

Glacigenic deposits of comparable age are found in the Timeball Hill and Makganyene Diamictite Formations of southwest Africa, which contain tillites dated at 2224 $\pm$ 21 Ma and 2300 $\pm$ 100 Ma (Visser, 1981). Another site of early Proterozoic glaciation is western Australia, where diamictites dated 2300 $\pm$ 200 Ma occur in the Meteorite Bore Member of the Kungarra Formation (Trendall, 1981). In Asia, the Gangou tillite of the Bijawar Group crops out in a narrow belt in the central part of India. The glacigenic beds are not well-dated, but are estimated to be about 1815 Ma (Mathur, 1981). Thus, it appears that the glaciation on the North American craton was part of a more widespread episode.

Paleomagnetic data, although sketchy, suggest the proposed ice center lay at a paleolatitude of about 60^{0} in North America, a typical position for mid-latitude ice sheets. Reconstructions of ancient continental positions vary, some showing Australia near southern Africa, and others showing it close to North America. This controversy makes it difficult to resolve the extent and location of the ice sheets, but one possibility is that there were two centers of glaciation, one in North America, another in southern Africa, India and western Australia. Whether the two ice centers represent temperate continental or high-latitude ice sheets remains unclear at the present time. (Young, 1979).

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